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Ninetieth Birthday of Academician Boris Sergeevich Sokolov

Academician Boris Sergeevich Sokolov, an outstanding Russian geologist, stratigrapher, and paleontologist, who contributed significantly to the development of these scientific branches, the editor in chief and founder of the journal *Stratigraphy. Geological Correlation*, celebrated ninetieth birthday in April, 2004.

Boris Sergeevich was born on April 9, 1914 (March 28 by the Gregorian Calendar), in Vyshnii Volochok, Tver guberniya, to the family of a doctor. Since childhood he took a keen interest in every living thing, he collected herbarium and arranged at home something like "a study of a young naturalist." Boris Sergeevich was also interested in inanimate nature, especially in fossils and minerals. From the earliest years, he engrossed in reading books about great geographic discoveries and voyages. Being out of school in his native town in 1932, Boris Sergeevich entered the Leningrad University (the Faculty of Geology, Soil, and Geography), where his interests in field geology, geography, and biology met a kind attention of talent professors, M.E. Yanishevskiĭ one of them. With his advances in studying a wide spectrum of geological branches and his field investigations in the Russian Platform, Sokolov who graduated from the university in 1937 was invited as an assistant to the Department of Paleontology of the faculty. Until the spring of 1941, Sokolov delivered the course of historical geology, continued investigations of Carboniferous fossil corals and Quaternary stratigraphy of the Moscow syneclise, which he carried out being a student, and actively studied the regional geology of Central Asia.

Results of these versatile investigations were such impressive that in the spring of 1941, he was appointed one of the leaders of the Complex Expedition to western China. The primary task of the expedition, which was held under difficult climatic conditions of the Chinese Tien Shan and adjacent deserts of Dzungaria and Tarim, was to study geology, ore and oil reserves of the Xinjiang (Sinkiang) Province despite a complicated political situation in China of those years. In that long-lasted expedition, Sokolov was accompanied by his wife Elena Nikolaevna Polenova and by little daughter Marina. Being a bright and surprisingly whole-hearted person, Elena Nikolaevna who became later an eminent specialist on ostracodes and Devonian biostratigraphy was devoted and nearest friend of Boris Sergeevich until she passed away in 1987. She created a unique warm atmosphere in their family and protected him against difficulties and troubles keeping the most favorable conditions for research of her husband.

The two-year expedition to China was successful, but the scientific report was not completed until 1950 and remained unpublished for a quite understandable reason, being deposited. Some information about the expedition appeared later in one of the essays written by Sokolov. In 1943, when the family returned to the USSR, Sokolov was invited immediately to join the Special Central Asia Oil Expedition of the All-Russia Research Geological Prospecting Institute (VNIGRI). The geological survey and assessment of oil reserves in areas adjacent to China have become a matter of his great concern for many years, especially for the period of 1943–1950, when he studied oil-and-gas fields in China, Kazakhstan, and Central Asia with a group of experts. Nevertheless, his main preoccupation was sedimentary cover of the Russian Platform, which was penetrated for the first time by deep reference boreholes down to the crystalline basement in a vast area from Baltic regions to the Urals and from the Arctic Ocean coast to the Black Sea.

On return to Leningrad at the end of 1945, Sokolov resumed his teaching and research activity at the university, continuing simultaneously his work on the oil geology at the VNIGRI. Sokolov's work in the post-war period is remarkable for diversity of scientific interests and comprehensive study of new materials. The most important lines in his work of that period were generalization of the Russian Platform paleogeography based on materials of the deep reference-drilling program just started and the monographic description of Paleozoic corals, especially of chaetetids and tabulates. Chaetetids are the fossil group, to which Sokolov devoted his dissertation (1947) finished prior to his departure to China in 1941. A five-volume monograph *Paleozoic Tabulate Corals in the European Part of the USSR* (1951–1955) became a true event in paleontology of corals. The monograph has rightly won international recognition as a classic paleontological work. A unified systematics of tabulatemorphic corals (including chaetetids assigned at that time to hydroids) was elaborated in this work for the first time, basic stages of their evolution were reconstructed, and stratigraphic implications were refined. In 1955, the monograph already published was successfully defended as his doctoral dissertation.

When analyzing materials of the deep drilling program in the Russian Platform during those years, Sokolov arrived at the conclusion that a thick complex of sediments existed there below the Cambrian base. This work resulted in distinguishing the mentioned complex as the first Precambrian System referred to by Sokolov as the Vendian. It also had a great effect on development of fundamentals of the Precambrian paleontology, stratigraphy, and biosphere. Sokolov was first to assess importance of paleobiological study in this part of geological records for development of biostratigraphy, for understanding the early stages of life evolution on the Earth, and for the doctrine of biosphere in general. In fact, the Precambrian paleontology, one of the problems of high priority, was under active elaboration since that time.

In 1958, Sokolov was elected to the USSR Academy of Sciences as corresponding member and moved to Novosibirsk, where organization of the Siberian Division of the academy was initiated at that time. Being a member of group, which included many prominent representatives of the Russian science, the leaders of that immense project, Sokolov headed the Department of Paleontology and Stratigraphy at the Institute of Geology and Geophysics. The department with dozens of young and talented specialists in paleontology, stratigraphy, and regional geology soon has gained international recognition for pioneering works. In those years, they were involved into comprehensive study within a broad stratigraphic range (Riphean–Phanerozoic) of the Siberian Platform sedimentary cover and surrounding structures. Still being busy with his investigation of corals (new monographs and a new volume *Fundamentals of Paleontology*), Sokolov was deeply involved into study of the Precambrian. The Vendian period in the history of biosphere became the main subject of his research. Established first in the Russian Platform based on geological data with due regard for problematic fossils of the genus *Beltanelloides* and microfossils, the Vendian substantiated by Sokolov, his disciples and followers as the terminal system of the Precambrian was acknowledged in this country and abroad. For his outstanding scientific and scientific–organizational achievements of that period, Sokolov was elected to the USSR Academy of Sciences in 1968 as its full member.

High professionalism and personal traits of Boris Sergeevich favored his versatile scientific–organizational activity, which was growing with time. Since the foundation of the Interdepartmental Stratigraphic Committee (ISC) in 1955, Sokolov was a member of its bureau, the chairman of the ISC Commission on the Ordovician and Silurian, the bureau member of several other commissions. He was vice-chairman of the ISC since 1971, the chairman (after Academician D.V. Nalivkin passed away) since 1976, and the honorary chairman since 1991. The All-Union (at present All-Russia) Paleontological Society (ARPS) was, and remains, of his special concern: in 1962 he was vice-president and since 1974 is president of the society. Every year he opens the ARPS sessions by delivering a special introductive report, in which timely problems and the most important achievements in paleontology are considered, and each his report is a significant scientific event for the audience. As scientist of the world-wide recognition, Sokolov became a member of the International Paleontological Union in 1964; he was vice-president of the Asian Division (1964–1966) and vice-president of the union (1970–1980). In 1980–1984, he acted as president of the International Paleontological Association. At the International Geological Congress of 1989 (Washington, USA), Sokolov was elected to the International Association of Coralists as the honorary president-organizer. He is also a member or honorary member of ten foreign academies and scientific societies. From the very beginning, Sokolov was involved into the International Geological Correlation Program (IGCP), one of the most efficient scientific programs of UNESCO.

A distinctive period in the life of Boris Sergeevich was in 1975–1990, when he entered Presidium of the Academy of Sciences in Moscow, and the general meeting of the academy elected him as academician secretary of the Division of Geology, Geophysics, Geochemistry, and Mining Science. During those years, a new strategy and mode of work formed in the division. Everybody bears in mind yearly reports by Sokolov at the general meetings of the division, in which he deeply analyzed achievements in basic fields of the Earth sciences, major aspects of applied geology,

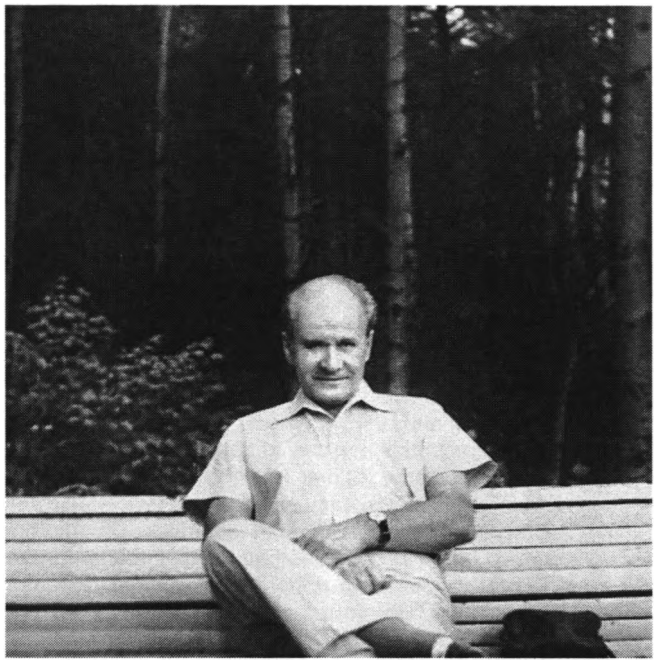
the state of mineral base and geological survey in this country.

Some other events are also related to Sokolov's moving from Novosibirsk to Moscow. In 1977, he organized for the first time in the country and headed the Laboratory of Precambrian Organisms (now the Laboratory of Precambrian Paleontology) at the Institute of Paleontology of the Academy of Sciences. One of the paramount achievements of laboratory was geological and paleontological substantiation of the Vendian System. The work involved leading specialists from Moscow, Leningrad, Novosibirsk, Kiev, and other cities. The two-volume monograph *The Vendian System: Historical–Geological and Paleontological Substantiation* (1985) edited by Sokolov was accepted worldwide, translated into English and issued abroad in 1992. These and other investigations of the laboratory founded by Sokolov soon won the general recognition and stimulated a series of major Russian and international projects on paleobiology and Precambrian stratigraphy. A sign of high appraisal of advances made by Sokolov and his colleagues is support of his scientific school by the Russian Foundation for Basic Research and by the Ministry of Industry, Science, and Technology of the Russian Federation.

Inherent of Sokolov, the science organizer, are firmness of purpose, exactness, and ability to foresee, define and clearly formulate the most topical lines of research. Also characteristic of him is aspiration to support young scientists and his ability to unite and inspire different specialists for solving the problems that define the foreground of science at present. Dozens of candidate and doctoral dissertations have been defended under his supervision.

In the Moscow period of life, Sokolov contributed much to the history of science, which was always the subject of his keen interest. Owing to his research and literary gift, readers learned a lot of new about life and creative work of distinguished geologists, paleontologists, and geographers of Russia and other countries. Moreover, some names were recovered by him and brought to public notice. For instance, an extensive article published in the *Pravda* has played a part in rehabilitation of V.I. Vernadsky. Even in a hard time of totalitarian censorship, he found possibility to publish works about prominent scientists who were not popular amidst the Soviet government, for example, Vernadsky, Lyubishchev, Teilhard de Chardin, and others. A great erudition and courage were necessary for this activity. Effective public actions by Sokolov for purity of the Lake Baikal and solution of other ecological problems are well known and live in our memory, like his other actions for strategy of science and education development in this country.

In addition to papers about science history, Sokolov published recently a series of heartfelt essays in the field that he refers to as "the family study of local lore." These publications triggered a full-scale



movement in the cultural life of central Russia. Important here were not only a great erudition and bright literary gift of Boris Sergeevich, but also his summer trips to the native village "Berezki" in the Tver oblast, where house of Academician Sokolov became an attractive center for intellectuals from Tver and St. Petersburg.

Scientific and scientific–organizational merits of Boris Sergeevich got worthy tributes: three Orders of Lenin, Order of the Red Banner of Labor, the Badge of Honor, and many medals. Sokolov was awarded the title of Hero of Socialist Labor in 1984, the Lenin Prize in 1967, Karpinsky Gold Medal (the highest academic tribute in the field of Earth sciences) in 1979, Karpinsky–Schweitzer Prize of the Hamburg Foundation, and Karl von Baer Medal for investigations in the Baltic Sea region in 1992. Academician Sokolov is author of more than 400 scientific works (fourteen books included) devoted to different aspects of geology, paleontology, stratigraphy, and general evolution of biosphere. In 1997, he was awarded the most prestigious tribute of the Russian Academy of Sciences, Lomonosov Grand Gold Medal, for his outstanding advances in studying the early biosphere of the Earth, distinguishing of the Vendian System, and classical works on fossil corals.

Sokolov's personal traits, his deep principles, benevolence, culture, unimpeachable self-control, democracy, approachability, personal interest in advances and fates of colleagues play a significant part in all the spheres of his successful work. Because of his deep cordial tactfulness, wonderful sense of humor, and marvelous fascination any personal contact with Boris Sergeevich is not only pleasant and helpful but inimitable as well.

The scope of scientific achievements, sophisticated investigations, striking literary gift, hard scientific–organizational work, and large school of followers hold Sokolov among the most brilliant representatives of Russian science. We wholeheartedly congratulate Boris Sergeevich with his ninetieth birthday and wish him the good health and new discoveries for the sake of science and motherland.

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Trends in Global Climate Changes Inferred from Geological Data

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Abstract—Recent paleoclimatic data reveal the following trends in climate changes. (1) During three billion years, characteristic of the Earth was gradual global cooling with the increasing frequency, duration and scale of glaciations. Based on these features, three principal climatic stages can be defined in geological history: (a) non-glacial (Early Archean), (b) with episodic glaciations (Late Archean–Middle Riphean), and (c) with frequent periodical glaciations (Riphean–Recent). (2) Irreversible climate changes were complicated and disguised by numerous superimposed temperature fluctuations of different periodicity and amplitude. In the Phanerozoic, a hierarchy of subordinate climatic fluctuations of 10–12 ranks, from extremely long (few hundreds million years) to short-term (tens years long only) is defined. Signs of climatic fluctuations of two–three highest ranks are recognizable in Proterozoic glacial sections. (3) The hierarchy of climatic fluctuations was stable during the Phanerozoic at least. (4) Amplitudes of climatic fluctuations depended on the cophasing degree of elementary climatic oscillations and character of their feedbacks in the biosphere. (5) The warm non-glacial climate prevailed during the Precambrian and Phanerozoic and was characteristic of 90% of the Phanerozoic and 95% of the post-Archean geological history. (6) Many climatic fluctuations, all those of first rank included, were of a global scale, synchronous, and cophasal. (7) Regional climate changes were caused by paleogeographic factors. (8) Global climate changes resulted in transformation of the latitudinal climatic zonality. The notion "global climate" is introduced to characterize the type of a planetary climatic zonality. Macrogeographic factors transformed latitudinal climatic belts into sublatitudinal ones. (9) Two main types of global climate (non-glacial and glacial) are defined. Transitions from non-glacial to glacial climate and vice versa were accompanied by rapid qualitative zonality reorganizations. (10) Each type of global climate is subdivided into gradations. (11) A peculiar feature of the global climate was an asymmetric position of climatic belts relative the equator. The asymmetry, which was insignificant during the non-glacial periods, and substantially increased at the time of glaciations, particularly of the great ones.

Key words: irreversible climate changes, quasi-periodic fluctuations, their hierarchy, global, synchronous, and cophasal patterns, climatic zonality, its reorganizations, asymmetry.

INTRODUCTION

General features of the Phanerozoic climate evolution became understood in the second half of the last century (Schwartzbach, 1974; Monin and Shishkov, 1979; and others) when several hypotheses were proposed to explain the climate changes (Monin and Shishkov, 1979; Budyko, 1980; Frakes *et al.*, 1992; and others). During the last decade, factual and methodological bases of paleoclimatology widened significantly and underwent qualitative changes. It was a consequence of research in difficultly accessible regions, deep-sea drilling, and of wide application of new approaches and methods (paleogeographic reconstructions based on plate-tectonic concept, geochemistry of stable isotopes, bed-by-bed study of sections using different methods, computer simulation of climate models, multivariant mathematic analysis of paleobotanic and lithological data, and so on). The quantity and quality of paleoclimatic data substantially increased. Accordingly, we got an opportunity to outline some general trends in climatic history of the Earth. The trends of empirical character are best detectable in the Phanerozoic history, although some of them are recognizable also in the Proterozoic, particularly in its late part. Some trends are distinct, while the others need in a further confirmation and should be considered as working hypotheses.

Particularly plentiful are data on the history and dynamics of glaciations, salt accumulation, stable isotopes, sedimentation settings, and paleobiogeography. These materials allow qualitative, semiquantitative, and, less commonly, quantitative estimates of climate temperature and humidity. The main attention is paid in this article to temperatures, because this leading factor is responsible for the regime of the Earth climate system and determines other climatic parameters, scale and intensity of global heat and moisture transfer inclusive.

1. DYNAMICS OF CLIMATE CHANGES

Well recognizable in the geological retrospective are irreversible climate changes and superimposed quasi-periodic climate fluctuations.

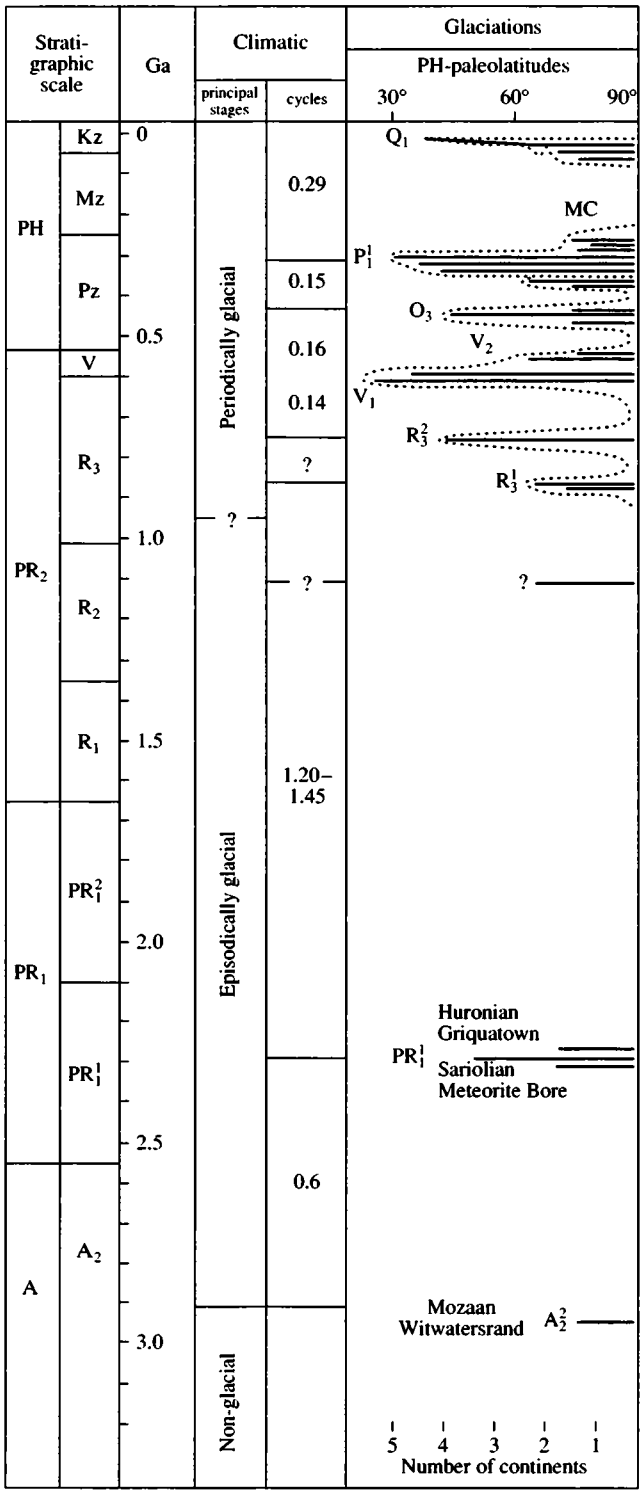

Fig. 1. Distribution of glaciations throughout the geological history; length of thin solid lines designating the glacial epochs is proportional to maximum distribution of glaciations (to paleolatitudes of the Phanerozoic and to number of continents in the Precambrian) during the given time, and dotted line envelopes glacial epochs of the Archean (A) and Mesozoic (MC) glacial periods.

1.1. Irreversible climate changes. The general trend in climate changes on the Earth can be inferred from glaciations. Their frequency and scale increased generally with time (Fig. 1), and this general trend cannot be a result of insufficient knowledge of older rocks only. During the last two–three decades, detailed geological mapping and prospecting covered almost the entire land areas composed of old rocks, those in developing countries included. Numerous and different new mineral deposits were discovered. The detailed studies could not miss glacial sediments, which form usually large bodies, are characterized by regional distribution, and attract attention of geologists by their unordinary appearance and origin. They showed that glaciations increased in scale during the geological history and their time structure was complicating. In addition, the studies of the last 30–40 years substantially specified ages and distribution areas of glacial sediments, although their new stratigraphic levels have not been discovered. An assumption based on astronomical hypothesis that the Earth glaciations started as long ago as in the Early Archean and repeated, beginning from 3460 Ma ago, every 400–370 m.y. (Steiner, 1978) has not been confirmed.[1] There is no also geological evidence of durable global glaciation on the early Earth, which was assumed by many researches in connection with the idea of a "weak" early sun. The planet surface was likely so "cool" in the Early Archean (Valley *et al.*, 2002) that its temperature was below 100°C, i.e., suitable for existence of hydrosphere at the surface.

The aforesaid allows a conclusion that available data reflect, with an appropriate approximation, a real distribution of glaciations in geological history.

The reliable signs of Early Archean glaciations are unknown so far. Their first signs, though scarce and spatially restricted, are established in the Upper Archean Witwatersrand Supergroup (Hambrey *et al.*, 1981) and Mozaan Group (Young *et al.*, 1998) of the relatively small Kaapvaal craton of South Africa. These glaciations are estimated to be approximately 2.9 Ga old (Nelson *et al.*, 1999). The Witwatersrand glaciation was probably of the piedmont or mountainous types while the Mozaan glaciation was of the ice sheet type because the last group encloses abundant dropstones (signs of ice rafting).

The next glaciations occurred only in the initial Early Proterozoic, i.e., 600–700 m.y. later than the Late Archean glaciations. The Early Proterozoic glaciations were of a substantially larger scale and mostly of the ice sheet type. Sediments they left behind are discovered on four remote continental blocks: on the Canadian Shield (Young, 1970), in South Africa (Visser, 1981), on the Baltic shield (Marmo and Ojakangas, 1984), and

[1] I use in this article the Russian stratigraphic terminology and Russian Precambrian geochronometric scale (Semikhatov *et al.*, 1991; Semikhatov, 2000; *Reshenie III Vserossiiskogo...*, 2001). For the Phanerozoic, the geochronometric scale by Harland *et al.* (1990) with some specifications for the Cambrian (Semikhatov, 2000) is used.

in western Australia (Martin, 1999; Eriksson *et al.*, 1999). Previously, I have considered these and the Witwatersrand glaciations as characterizing one into the Canadian Glacioera of the Early Proterozoic age (Chumakov, 1978). Now, it is established that the Witwatersrand (the Kaapvaalian, in general) glacial sediments are significantly older, Archean in age, whereas the Lower Proterozoic glacial sediments proper are estimated, based on generalized radioisotope data, to be within the narrow interval of 2.3 to 2.2 Ga (Crowell, 1999). Therefore, the Kaapvaalian and Early Proterozoic glacial events should be regarded separately as corresponding to the Kaapvaalian and Canadian glaciation periods, but not to one Canadian Glacioera. Similar C-isotope anomalies in carbonate rocks overlying sediments of the Canadian glacial period confirm their approximate synchronism (Semikhatov *et al.*, 1999; Bekker *et al.*, 2001). In North America, the Canadian glacial sediments are known from several remote regions. In the Great Lakes area, the Huron Supergroup encloses sediments of three large glacial epochs separated by interglacial deposits. In turn, the Late Gowgandan glacial epoch comprised two glacial events (Young, 1970).

Glaciation indications were never discovered at the higher levels of the Lower Proterozoic and in the Lower and most Middle Riphean sections (Fig. 1). The lack of large C-isotope anomalies within the stratigraphic interval of 1000–1200 m.y. long that is termed sometimes as the "glacial pause" indicates likely the absence of glaciations (Bekker *et al.*, 2001). According to brief descriptions published (Salop, 1973; Akhmedov, 2001; and others), scarce scattered boulders in the upper Lower Proterozoic sections are most likely of the landslide and volcanogenic origin or dispersed by seasonal ice, being indicative of the moderately cold though non-glacial climate.[2] Some researchers assumed that the "glacial pause" was caused by the elevated content of methane in the atmosphere, whereas glaciations, which bounded the "pause," were to the contrary related to episodes of atmospheric methane oxygenation (Pavlov *et al.*, 2003). Nevertheless, a wide distribution of red beds (Eriksson *et al.*, 1992; and others) and even tropical laterites (Beukes *et al.*, 2002) in this stratigraphic interval, which indicate the existence of oxidizing atmosphere at corresponding time, is inconsistent with this hypothesis.

It is unclear so far weather glaciations existed in the Middle Riphean or not. In the Baikal–Patom Highland and Brazil, there are glacial sediments, which can be as old as the Middle or Late Riphean (Chumakov, 1993a; Khomentovskiĭ and Postnikov, 2001). Beginning from the Late Riphean until Recent, glaciations occurred on the Erath regularly, and their scale increased with time. During their maximums, the glaciations spread over large territories (4–5 continents at once) to the latitude of 40°–30° and, sometimes, even to lower latitudes.[3] The glacial periods became complicated, consisting of multiply repeated glacial epochs and subordinate smaller-scale glacial events.

Thus, during the last three billion of years, the increasing role of glaciations characterized the main trend in climate changes on the Earth. This indicates a gradual cooling of the planetary surface. During the first two billion of years, glaciations were rare. The total duration of glacial periods corresponds to 7%, not more, of this time interval. Approximately 1 Ga ago, cooling was substantially intensified, and glacial periods in total lasted longer, approximately 25% of one billion of years long interval and 30% of the Phanerozoic history, i.e., during the last 535 Ma (table).

Judging from the facts and considerations mentioned above, three principal stages are recognizable in the Earth history: (1) *non-glacial* (Early Archean); (2) *with rare episodic glaciations* (Late Archean–Middle Riphean); and (3) *with frequent periodic glaciations* (Late or Middle Riphean to present time).

The main factors responsible for global cooling were probably the changes in atmosphere composition, particularly the CO_2 content decrease. This could be related to the lowering intensity of volcanism (Abbote and Isley, 2002a) and endogenic degassing of the Earth, on the one hand, and to intensified burial of organogenic carbon and carbonates in the sedimentary shell, lithosphere, and mantle in the course of plate tectonics development, on the other. This is evident from correlation between the principal climatic stages mentioned above and main stages in evolution of plate tectonics (Chumakov, 2001b). The intensity of carbon burial increased with development of weathering processes and photosynthesis. Global cooling could also be stimulated by the planetary albedo increase with the growth of continents. Reduction of endogenic heat flow could also contribute to cooling, although this process was subordinate, because the endogenic heat proportion in the Earth's heat balance was by two–three orders lower than the heat contribution produced by insolation.

1.2. Periodical climate changes. The general trend of the Earth surface cooling was not unidirectional. It was complicated and disguised by numerous climate changes of different signs and scales. Most of these changes occurred within certain time limits, and the state close to the initial one returned some time later. Therefore, all or almost all climate changes can be con-

[2] If it appears that some shales with boulders occurring in the upper part of the Lower Proterozoic section are of glacial origin, this will not change the conclusion that the frequency and scale of glaciations gradually increased during the post-Archean interval of geological history.

[3] Nevertheless, the hypothesis of global glaciations in the Vendian and Late Riphean (Harland, 1964; Hoffman *et al.*, 1998; and others) cannot be considered as proven because it is still based on scarce (Evans, 2000) and doubtful (Meert and Van der Voo, 1995) paleomagnetic determinations and is poorly consistent with several geological facts (Chumakov, 1992, 2003) and with results of climate mathematical modeling (Poulsen, 2003; and others). Data in favor of the Early Proterozoic global glaciation (Evans *et al.*, 1997) are even less reliable.

Glacial and non-glacial climatic periods and associated events

Age, Ma	Glacial (g[1]) and non-glacial (t)	Duration, m.y.	Polar continents[3]	Super continents[2]	CO_2 content, n[4]	Maximums		
						tectogenesis[5]	volcanism	
							subduction-related	mantle
0								
Kz_{2-1}	ggggg	40	++	+	1	+	+	
40			+					
Kz_{2-1}–Mz	t	205	++	++	5–7			++
245						++		
Pz_3^2	ggggg	78	++	++	1	+		
323								
C_1	t	30	++	+	2–4			
353								
C_1–D_3	ggggg	14	++	+	4–7		+	
367								
D_3–S_3	t	59	+	+	7–12	++		+
426								
S_2–O_3	ggggg	18	+	+	12–17	+	+	
444								
O_3–Cm_1	t	86	+	+				
530								
Cm_1–V_2	gg	10 ?	+	+				
540								
V_2	t	45	+					+
585								
V_1	ggggg	15 ?	+	+				
600								
R_3^{2-3}	t	140						+
740								
R_3^2	gggg	20 ?		+				
760								
R_3^{1-2}	t	130		++				+
890								
R_3^1	ggg	20 ?						
910								
R_{3-2}^1–PR_1^2	t	1290						+
2200								
PR_1^2	gggg	100						
2300								
PR_1^1–AR_3	t	600						+
2900								
Ar_3^2	ggg	50 ?						
2950								

[1] the number of letters designates approximate glaciation scale; [2] Gondwana-type (+) and Pangea-type (++) supercontinents; [3] glaciations on one (+) and two (++) poles; [4] n present-day CO_2 content in the atmosphere; [5] significant (+) and maximal final (++) phases.

sidered at first approximation as climatic fluctuations of different periodicity. The periodicity notion is used here *sensu lato* as it is usually understood in geology. In reality, these fluctuations were not strictly periodical, harmonic, and, consequently, linear. This is evident from the patterns of climatic secular successions and from bifurcations of oscillation periods inferred from their spectrum–time analysis (Chumakov and Oleinik, 2002). Strictly speaking, the fluctuations should be termed as self-similar or similar. Periodic fluctuations are sufficiently well established in the Phanerozoic and, to a lesser extent, for the Vendian and Late Riphean; sometimes, they are recognizable in older intervals as well. The spectrum of periodic climatic fluctuations was very wide: from several tens years to hundreds million years. Different-period fluctuations were superimposed on each other (Fig. 2) and, as a rule, non-harmonic. These features and approximate character of many geochronological dates hamper the exact determination of the fluctuation periods duration. Nevertheless, oscillations of several ranks and their groups are usually well distinguishable in paleoclimatic curves (Douglas and Woodruff, 1981; Zakharov, 1992; Chumakov, 1993b; Fot'janova and Serova, 1994; and many others) as well as in spectrum (Imbrie *et al.*, 1984; and others) and spectrum–time (Chumakov and Oleinik, 2002) diagrams. This implies that many fluctuations substantially differed from each other in duration. They are divisible into groups differing from each other in fluctuations average duration that varied several-fold and, sometimes, by an order of magnitude. Let us consider periodical climatic fluctuations, which are recognizable by geological methods, beginning with the large-scale ones. It is convenient to distinguish fluctuations of the following five groups.

Superlong climatic fluctuations. The idea of a large-scale climatic periodicity was formulated long ago. Holmes (1937), Umbgrove (1947), and Lungershausen (1956) were among the first to propose such an idea based on distribution of glaciations through geological history. Because of poor knowledge of ancient glaciations and insufficient isotope dating, the suggested periodicity was largely intuitive. These authors estimated the duration of climatic (glacial) cycles as ranging from 250 to 190–200 m.y. The other estimates published afterward are 300 (Keller, 1972), 300–1200 (Avdeev, 1973), 280–400 (Steiner, 1978), 217 (Zakoldaev, 1991), and 215 m.y. (Yasamanov, 1993). Assessments of Yasamanov and some other researchers are based on incomplete and partly outdated paleoclimatic and geochronological data. In addition, these data were selected and interpreted rather *ad arbitrum* leaning upon the duration of the current galactic year or deductively calculated duration of past galactic years. It should be noted that astronomers estimate very approximately (from 180 to 300 m.y.) even the duration of the current galactic year. It is more reasonable therefore to define the large-scale climatic periodicity using an inductive approach based on trustful paleoclimaic data. Estimating time intervals between dates most frequently mentioned for the Late Precambrian and Phanerozoic glaciations, Williams (1975) arrived at the conclusion that climatic cycles were approximately 150 m.y. long. Afterward, Frakes *et al.* (1992) estimated that time spans between the middles of Phanerozoic and Late Precambrian cold and warm periods range from 138 to 181 m.y. Based on the Phanerozoic curve of secular $\delta^{18}O$ variations, some researchers suggested that prevalent climatic cycles lasted approximately 135 ± 9 m.y. (Shaviv and Veizer, 2003).

Because of significant scatter of estimated duration of climatic cycles, some authoritative researchers cast doubts on existence of any large-scale climatic periodicity at all (Harland, 1981; House, 1995; Crowell, 1999). Their skepticism is not quite sound. It was already mentioned why the deduced climatic cycles are of discordant duration. As for the empirical approach, discrepancies are related, largely, to the fact that transitions from warm to cold intervals and vice versa were rather gradual and lasted tens of million years. In addition, transitional periods were often poorly studied and duration of cold and warm epochs was estimated very arbitrarily, i.e., subjectively. When periodicity is established based on most prominent and accurately dated climatic events, par example, on glacial maximums, the discordance can be reduced substantially. Let us consider this problem in more detail. There are known five glacial maximums of the Late Riphean–Paleozoic time termed sometimes as "great glaciations": (1) first Late Riphean event of this kind was 850–890 Ma ago, (2) second Late Riphean 740–750 Ma ago, (3) the Early Vendian 600 Ma ago, (4) Late Ordovician, 440 Ma and (5) Late Paleozoic 290 Ma ago. The last four "great glaciations" are separated by time intervals of 140, 160, and 150 m.y., respectively. The method substantially decreases the scatter of cycles duration. In fact, it appears to be within accuracy limits of stratigraphic methods. Accordingly, the existence of a regular climatic periodicity within the Late Riphean–Paleozoic interval quite apparent (Chumakov, 2001a). It is probable that the interval between the first and second Late Riphean glaciations was also as long as 140–150 m.y. The last Paleozoic (Asselian–early Sakmarian) glacial maximum is separated from the next (Pleistocene) one by the time interval of 290 m.y. If we take into consideration the fact that signs of cooling and, locally, even of seasonal ice rafting are recorded in the terminal Jurassic–initial Cretaceous, i.e., in the middle of this interval (Hambrey *et al.*, 1981; Frakes *et al.*, 1992), then two additional Mesozoic–Cenozoic climatic cycles each 140–150 m.y. long can be outlined. These five or, probably, six climatic cycles lasting 150 m.y. are close in average duration to those established by Williams and Frakes with colleagues. Periods of similar duration are inferred also from the spectrum and spectrum–time analyses of semiquantitative paleoclimatic data available for the Late Riphean and Phanerozoic (Chumakov and Oleinik, 2002). The duration value

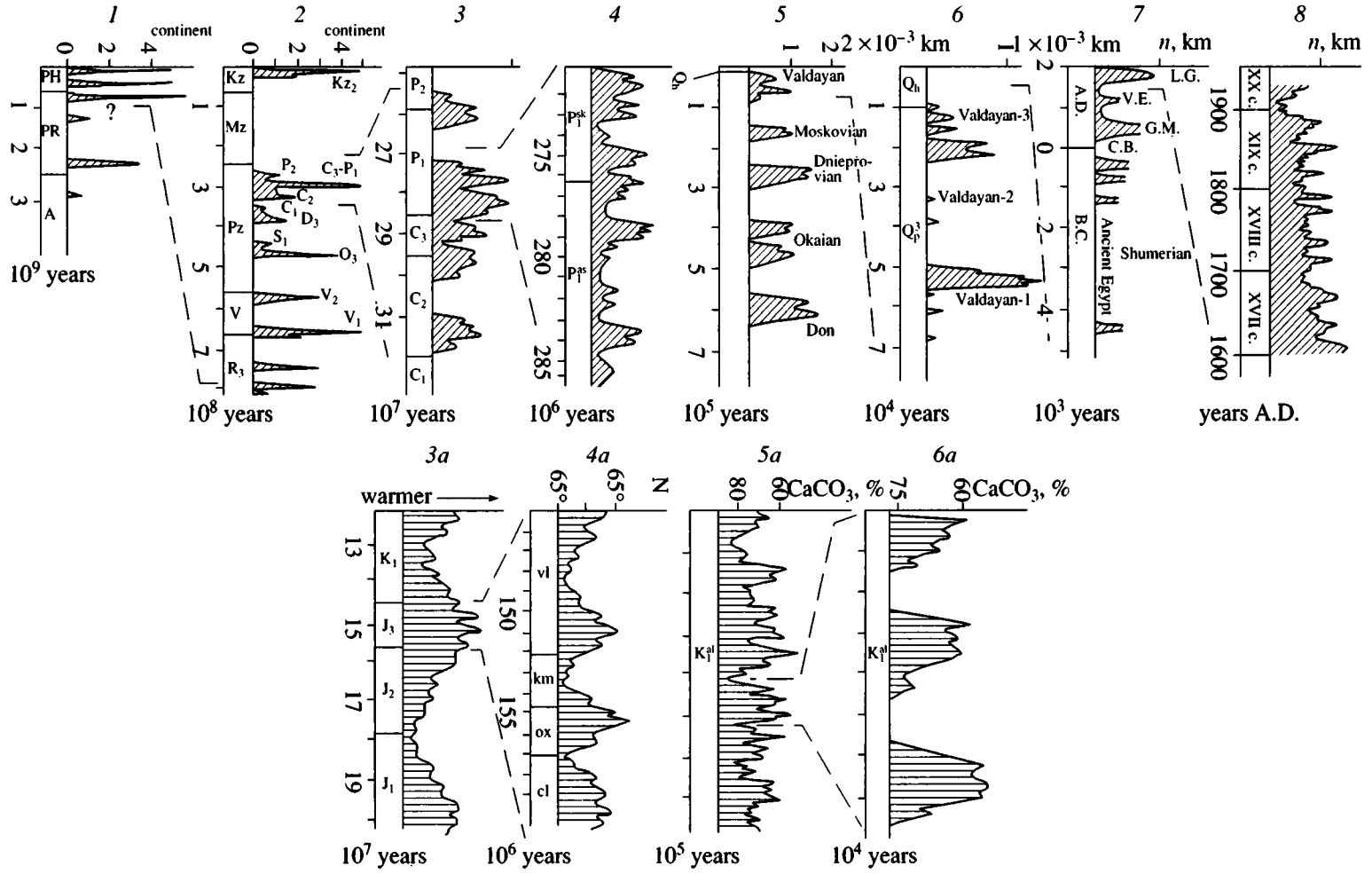

Fig. 2. Upper diagrams exemplify glacial events of different ranks (hatched) and relevant climatic fluctuations (Chumakov, 1993b with additions); lower diagrams illustrate climatic events and fluctuations of similar ranks in non-glacial intervals of the Phanerozoic (remember 10 times difference between scales of adjacent diagrams): (*1*) three principal climatic stages AR_1, AR_2–R_2, and R_3–Recent; (*2*) glacial and non-glacial periods, superlong climatic fluctuations; (*3*) glacial and non-glacial epochs, long climatic fluctuations, C_2–P_2, southeastern Australia; (*3a*) warmer and cooler epochs (thermal and semithermal epochs), J–K_1, Siberia (after Zakharov, 1992); (*4*) glacial and non-glacial ages, middle-ranged climatic fluctuations, P_1^{as}–P_1^{sk}, western Australia; (*4a*) warmer and cooler ages (thermal and semithermal ages), J_3, and Callovian (cl), Oxfordian (ox), Kimmeridgian (km), Volgian (vl) invasions of Tethyan fauna into basins of northeastern Asia (after Velichko *et al.*, 1994); (*5*) glacials and interglacials, short climatic fluctuations, Pleistocene, East European Plain; (*5a*) short (Milankovitch-type) fluctuations reflected in color variations and $CaCO_3$ contents in sediments of the *Ticinella praeticinensis* Subzone, K_1^{al}, Italy (after Larson *et al.*, 1993); (*6*) stadials and interstadials; short climatic fluctuations; Q_p^3, East European Plain; (*6a*) the same as *5a*, but of the lower rank; (*7*) secular fluctuations of present-day glaciers, Q_h, Tirol, (L.G.) "Little Glacial Age," (V.E.) Vikings Epoch, (A) Archean, (G. M.) great migration of people; (C. B.) Christ birth; (*8*) shorter fluctuations of present-day glaciers (17th to 20th centuries), Alps.

inferred from the Phanerozoic $\delta^{18}O$ curve (approximately 135 ± 9 m.y.; Shaviv and Veizer, 2003) is rather close to values mentioned above.

In addition to prominent cycles 140–150 m.y. long, there are also even longer, although less manifested cycles. Every second "great glaciation," beginning from the Vendian one, is of a slightly larger scale. Therefore, the largest glaciations (Pleistocene, Late Paleozoic, and Vendian) form two cycles approximately 300 m.y. long. The climatic cycles 140–150 and 300 m.y. long can be united, for convenience, into a group of superlong fluctuations (Chumakov, 1995a). They determined main climatic changes during the Phanerozoic, Vendian and Late Riphean.

The causes of superlong climatic fluctuations were discussed for a long time by many experts. The total number of published hypotheses about causes of glacial periods is close to a hundred (for critical review of most important hypotheses see Chumakov, 2002). Here, I consider briefly only some of them. As was noted, many researchers attempted before and are trying now to correlate superlong climatic cycles with the galactic year (Umbgrove, 1947; Lunsgerhausen, 1956; Keller, 1972; Zakoldaev, 1991; Yasamanov, 1993; and others) or with the galactic half hear (Williams, 1975; Frakes *et al.*, 1992). The cycles 140–150 m.y. long are however substantially shorter as compared with both synodic and anomalistic galactic years, whereas the comparison of glacial cycles with the galactic half year is illogical, unless additional complex assumptions are introduced (Williams, 1975; House, 1995).

Quite widespread is an idea that alternation of glacial and non-glacial periods was mainly determined by oceanic circulation (the latest example is publication by Smith and Pickering, 2003). This hypothesis does not take into account the fact of very close configurations of continents and oceanic currents at the commencement (Eocene, Antarctica) or termination (Permian, Gondwana) of some glacial periods.

The glacial periodicity correlates significantly better with the tectonic and volcanic activity of the Earth (Chumakov, 2001a). The glacial maximums were synchronous with early phases of tectonic cycles and peaks of subduction-related explosive volcanism. This leads to the conclusion that accumulation of thick sedimentary sequences during the early phases of tectonic cycles and associated intense weathering of silicates determined the burial of large volumes of carbonates and organic matter, reduced the CO_2 content in the atmosphere, and set the stage for glaciations (Lindsay and Brasier, 2002; Schrag *et al.*, 2002). The "volcanic winters" related to intensified explosive volcanism triggered glaciations. Owing to numerous and strong positive feedbacks in the biosphere, glaciations became stable and widened.

Glacial maximums predated the main final phases of tectonic cycles. During maximal phases of tectogenesis and orogeny, glaciations ceased or rapidly degraded. This can be explained by reduced subduction-related volcanism, increased atmosphere transparency, and elevated heat balance at the Earth surface. The processes that accompanied main orogenic phases, such as intense erosion, granite formation, and regional metamorphism in orogenic belts, resulted in oxidation of organic and carbonaceous matter in sedimentary sequences and in decomposition of carbonates in mixed terrigenous–carbonate deposits ("dedolomitization" and "decarbonatization"). Some periods were marked by the intensified magmatism of mantle plumes. The released carbon dioxide entered the atmosphere and stimulated additional warming. Being combined, these processes resulted in substantial warming and termination of glaciations.

It cannot be entirely ruled out that periodicity of tectonic and volcanic activity, and to a certain extent, the periodicity of global climate were controlled by cosmic (Abbot and Isley, 2002b) or galactic processes, for instance, by the solar system pass through sleeves of spiral star accumulations (Shaviv and Veizer, 2003).

"Great glaciations" corresponded to peaks of glacial epochs, duration of which varied in the Late Riphean–Phanerozoic from approximately 10–15 to 78 m.y. averaging approximately 27 m.y. The glacial epochs were separated by durable non-glacial periods (interglacials) that lasted in the Late Riphean–Phanerozoic from 30 to 205 m.y. averaging approximately 95 m.y. Climatic cycles corresponding to alternating glacial periods (Middle Mesozoic cooling included) and interglacial periods show no regularity and range from 55 to 180 m.y. Taking into consideration a rather strict periodicity of great glaciation maximums, irregularity of related climatic cycles of lower ranks represents a problem, which requires special discussion. Here, it should be noted only that this is probably caused by different combinations of paleogeographic, geochemical, and geodynamic factors that accompanied glaciations and influenced, to some extent, the latter.

Long climatic fluctuations. The abovementioned glacial periods consisted of separate shorter events: glacial and interglacial epochs, which lasted from a few to 10–15 m.y. (Figs. 1, 2). Biostratigraphic and geochronometric methods, which can be used to measure duration of glacial and non-glacial periods, are of insufficient resolution in many cases for determination of climatic epochs duration. The average duration of these epochs can approximately be estimated, when several of them are recognizable in a continuous sedimentary succession of determined time-range. For the Late Paleozoic, such a succession can be exemplified by the Hunter River valley section in eastern Australia, where glacial sediments repeatedly alternate with marine varieties (Chumakov, 1993b). Some researchers believe that basal layers of this succession are of the Namurian Stage. According to the other standpoint, the base of this glacial section is the Westphalian in age. The glacial section is crowned by the uppermost Kungurian or

Kazanian sediments. There are indications that iceberg-rafted coarse debris accumulated in southeastern Australia until the Tatarian Age (Veevers *et al.*, 1994). Thus, the integral duration of the Carboniferous–Permian glacial period in this region is estimated as ranging from 55 to 67–70 m.y. according to the most widely accepted scale (Harland *et al.*, 1990). In the Hunter River section, recognizable are three glacial units (tillites, iceberg-related, fluvioglacial, and glacial-marine sediments) and two intervening interglacial units (terrigenous marine and coaliferous sequences), i.e., five climatic epochs in total. The average duration of these epochs ranged from 11 to 14 m.y. Thus, the duration of climatic cycles of two, glacial and interglacial epochs can be estimated as 20–30 m.y. long. When recent data on probable Namurian age of basal layers in the Carboniferous–Permian Dwyka Formation (Streel and Theron, 1999) are taken into consideration, the similar average duration of 9 m.y. appears to be characteristic of four glacial and three intervening non-glacial epochs in South Africa (Visser, 1997). The cycles formed by these epochs are as long as approximately 18 m.y. Climatic fluctuations with periods of a few tens million years long, although of a lesser amplitude, are also recorded in ice-free realms. During the Late Paleozoic glaciations, such fluctuations were recorded in central Siberia, where their duration is estimated to be about 35 m.y. long (Chumakov, 1995a). As for Cenozoic glaciations, recognizable are climatic fluctuations with periods of 15 to 20 m.y. in northeastern Asia (Velichko *et al.*, 1994), of about 11 to 14 m.y. in Kamchatka and Sakhalin (Fot'janova and Serova, 1994), of 15 m.y. in Sakhalin and the Primor'e region (Krassilov, 1994), and in southern West Siberia and southern East Europe as well (Velichko *et al.*, 1994). Differences between the estimated durations can be explained by insufficient accuracy of available stratigraphic data and correlation of continental sediments, particularly of the Permian deposits.

Climatic fluctuations a few tens million years long were first united into a group of long-wave or long oscillations within glacial sections (Chumakov, 1993b), although they are characteristic also of non-glacial Phanerozoic periods. These fluctuations were manifested as alternation of more and relatively less warm climatic epochs. Taking into consideration the fact that differences between these epochs are relative only, they can be termed as thermal and semithermal epochs. Thermal and semithermal epochs are distinct in the Mesozoic of West Siberia (Gol'bert, 1987; Zakharov, 1992), in the Late Cretaceous of mountainous (Krasheninnikov *et al.*, 1990) and coastal (Herman, 1993) areas of northeastern Asia, in the Mesozoic–Cenozoic of low latitudes (Douglas and Woodruf, 1981), in the Falkland Plateau area of the Southern Hemisphere (Krasheninnikov *et al.*, 1990), in middle latitudes of the Indian Ocean (Clarke and Jenkyns, 1999), and in the Paleogene of Antarctica (Dingle and Levelle, 1998). For Siberia, their typical periods are estimated to be 11–35 m.y. long (Chumakov, 1995a) and their amplitudes as corresponding to several degrees (Gol'bert, 1987).

Like superlong climatic periods, the long-wave fluctuations were probably related to variations in intensity of tectonic and magmatic processes on the Earth (Chumakov, 2001a).

Middle climatic fluctuations. During the last decades, it was established that climatic epochs were also non-uniform and consisted of relatively cold and warm time intervals. Inasmuch as these intervals are subordinate to climatic epochs, it is logical to term them, according to accepted system, climatic ages: glacial and interglacial ages of glacial epochs, and semithermal and thermal ages of non-glacial epochs. Climatic ages were usually many hundreds of thousand to several million years long. Cycles of alternating cold and warm ages, which lasted from about a million to tens million years, were defined as middle-wave climatic fluctuations (Chumakov, 1993b). At present, there are numerous examples of such fluctuations characteristic of both the glacial and non-glacial climatic epochs (Fig. 2). In addition to the Late Ordovician, Silurian, Early Permian, Jurassic, and Late Cenozoic middle climatic fluctuations mentioned in previous publications (Chumakov, 1993b; Velichko *et al.*, 1994) they are established in many other regions. These fluctuations are 1.2 m.y. long in the Ordovician–Silurian of Arabia (Vaslet, 1990), about 1.3 m.y. long in the Silurian of South America (Grahn and Caputo, 1992), 2 to 5 m.y. long in the Aptian–Eocene deposits of Indian Ocean (Clarke and Jenkins, 1999; Chumakov and Oleinik, 2001), 4 to 10 m.y. long in the Late Cretaceous of the Far East (Zakharov *et al.*, 1999), 1.7 to 3.6 m.y. long in the southern Atlantic (Herbert *et al.*, 1999), 1.5 to 8 m.y. long in the Oligocene–Miocene worldwide (Zachos *et al.*, 2001), and 1 to 2.7 m.y. long in the Late Cenozoic of the Arctic (Velichko and Nechaev, 1999).

For a long time, researchers ignored the middle-wave climatic fluctuations flattening them in secular climatic curves and leaving aside by paleoclimatic reconstructions and climate mathematical modeling. Nonetheless, these fluctuations larger in scale than those of Milankovitch type prevail (over 50%) over the others (Chumakov, 1995a). They could be responsible for anomalous temperature oscillations of 10–12°C, as estimated for the Late Cretaceous of the Far East (Zakharov *et al.*, 1999), and of 3–6°C, as established for the Oligocene–Early Miocene deep oceanic sediments (Zachos *et al.*, 2001). Ice-rafting episodes in the Greenland and Norwegian seas were also related to the middle-wave climatic maximums (Velichko and Nechaev, 1999).

Causes of middle climatic fluctuations are unclear being under active debates. Some researchers argue that they are related to long-lasting variations in eccentricity of the Earth orbit (Herbert *et al.*, 1999). Remarkable is the proximity between periods of middle climatic fluc-

tuation and periods of high-frequency eustatic fluctuations.

Short climatic fluctuations. Short climatic fluctuations from a few tens to hundreds thousand years long (Milankovitch fluctuations, after Imbrie *et al.*, 1984) are known for a long time as characterizing alternating glacials and interglacials of the Pleistocene and their phases in high and middle latitudes. These fluctuations were of a global scale and manifested, although to a lesser extent, in low latitudes and in all subsystems of the biosphere as well. On land, short climatic fluctuation are readily recognizable in loess–soil (Zubakov, 1986; Dodonov, 2001) and lacustrine (Karabanov *et al.*, 2000, Lowenstein *et al.*, 1999) sequences being also inferable from alternation of humid and more or less arid environments and from respective vegetative communities (Dupont *et al.*, 2000; Van der Kaars and Dam, 1995). In seas and oceans, these fluctuations resulted in changes of surface (Velichko and Nechaev, 1999) and deep-water (Chapman and Shackleton, 1999) temperatures, sea-level oscillations, and substantial changes in sedimentation patterns (Lisitsyn, 1988). They are also responsible for changes in faunal assemblages (Barash *et al.*, 1989; and others), primary bioproductivity (Beaufort *et al.*, 1997), variations of oxygen (Imbrie *et al.*, 1984) and carbon isotope compositions, and for other sedimentological, biotic, and geochemical events. In the atmosphere, short climatic fluctuations were accompanied by changes in the carbon dioxide, hydrogen, and dust content, and in the oxygen isotope composition, as it is inferred from investigation of gas inclusions in ice core samples (Kotlyakov and Lorius, 2000).

It was believed for a long time that short climatic fluctuations are characteristic of glacial periods only (Woldstedt, 1954; Velichko, 1987). At present, it is shown that these fluctuations prevailed, in various forms, during the entire Phanerozoic at least, in its glacial and non-glacial periods. Listed below are only some of numerous publications, where short climatic fluctuations of various pre-Pleistocene subdivisions are mentioned: Pliocene (Zubakov and Borzenkova, 1983; Raymo, 1992; and many others); Miocene and Paleogene (Zachos *et al.*, 2001); Cretaceous (Larson *et al.*, 1993; Mutterlose and Ruffell, 1999); Jurassic (Waterhouse, 1999); Permian (Anderson and Dean, 1995), Carboniferous (Weedon and Read, 1995; Miller and Eriksson, 1999), Devonian (Wu *et al.*, 2001), Silurian–Ordovician (Williams, 1991); and Ordovician (Sutcliffe *et al.*, 2000).

After works by J. Adhemar, J. Croll, and, particularly, by M. Milankovitch, many geologists began to relate Pleistocene glacial events to variations in orbital parameters of the Earth, the inclination of rotation axis included. The study of oceanic sediments in the 1970s showed that the oxygen isotope composition variations in carbonates, which reflect the ice volume on the planet, are well correlative with astronomical periods of 19, 23, 41, and, particularly, 100 thousands years (Imbrie *et al.*, 1984). Variations in paleotemperatures and several other climatically important parameters with a close periodicity were revealed by geochemical study of ice cores from Greenland and Antarctica (Kotlyakov and Lorius, 2000). In addition, a cycle 400–410 thousand years long is recorded in most complete sections of both the glacial and non-glacial sediments (Herbert *et al.*, 1999). These data, as well as numerous age estimates obtained by isotopic and other methods confirm earnestly the astronomical control of short variations. The fact that short climatic fluctuations occurred through the entire Phanerozoic regardless of repeated changes in geological, geographic, climatic, and biotic situations on the Earth is an additional argument in favor of the astronomical control.

Ultrashort climatic fluctuations (millennial, centennial, and shorter). These fluctuations are registered by instrumental observations and historical, archeological, dendrometric, glaciological, and geological data on the Holocene and Late Pleistocene (Broecker and Denton, 1989; Bianchi and McCave, 1999; Lee and Slowey, 1999; Kotlyakov and Lorius, 2000; and many others). Some of these fluctuations show certain correlation with the solar activity. The radiocarbon analysis of wood growth rings established, in particular, the cycles of relatively elevated solar activity, which are approximately 2400, 299, and 90 years long. By duration and phases, the first two cycles are almost identical to the Holocene climatic cycles (Dergachev, 1994; Vasil'ev *et al.*, 2002; Raspopov *et al.*, 2000). In older sediments, ultrashort fluctuations, for instance those, which are 200 years long, are recognized only in some cases (Anderson and Dean, 1995).

Weather fluctuations. The climate is characterized by average perennial meteorological parameters. Their changes with periods below 40 years are usually referred to weather fluctuations. As is known, dominant among the latter are annual (seasonal) and daily fluctuations. In addition, there are perennial cycles, some of which are also related to variations in the solar activity. Sometimes, seasonal (Chambers *et al.*, 2000; and many others) and perennial climatic cycles are recognizable, with a certain confidence, in older sediments as well.

Summing all the mentioned data on climatic periodicity, we can state that the Phanerozoic was characterized by the hierarchy of climatic fluctuations of 10–12 ranks at least (Fig. 3).

2. CLIMATIC FLUCTUATIONS IN THE PRECAMBRIAN

Currently, the problem of climatic fluctuations in the Precambrian is of particular interest. Many geochemical, stratigraphic, and paleogeographic interpretations, as well as an objective estimate of the hypothesis of global glaciations are connected with this problem.

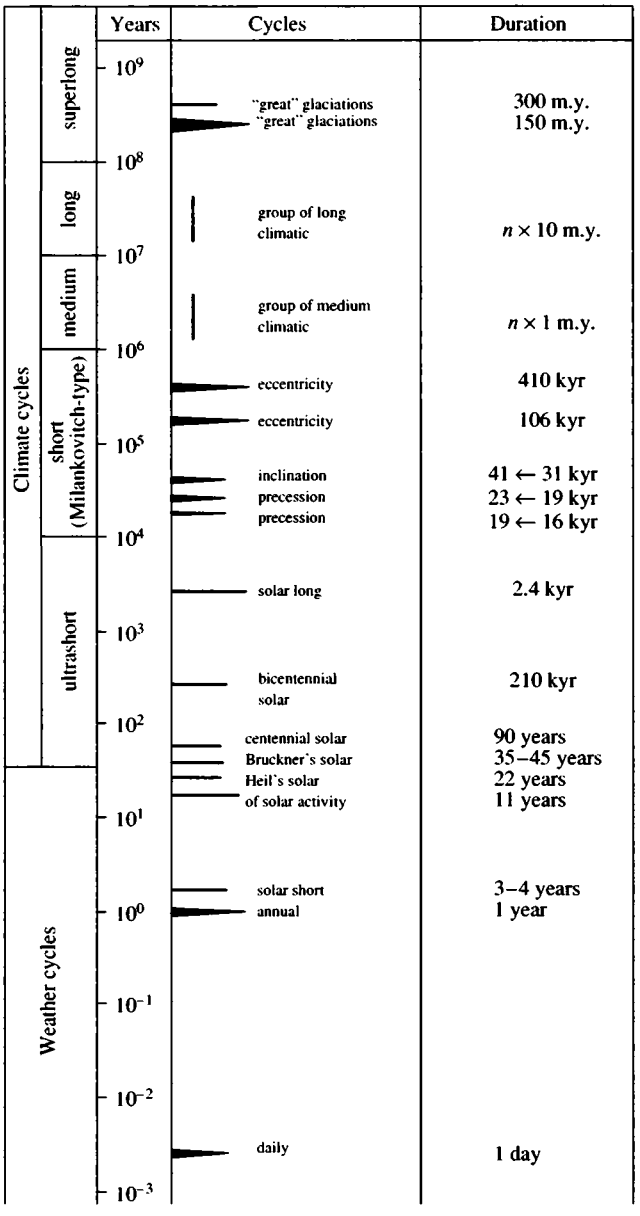

Fig. 3. Hierarchy of climate and weather cycles (thick solid lines designate the most important cycles).

The low resolution of biostratigraphic methods for this time interval and scarcity of reliable radiometric ages hamper usually a direct estimate of periodicity in Precambrian climatic fluctuations. Nevertheless, analogues of largest Phanerozoic climatic fluctuations can be recognized, with certain confidence degree, in the Proterozoic as well. Let us briefly consider some typical Precambrian glacial sections from the viewpoint of climatic fluctuations.

The Vendian and Riphean. The occurrence of two superlong climatic cycles approximately 140–150 m.y. long in the Vendian–Late Riphean is quite obvious (Fig. 1). Glacial and interglacial sediments alternating in most complete sections correspond to separate intervals of superlong Precambrian cycles and suggest existence of subordinate shorter cycles. By facies peculiarities, their combinations and thicknesses, such sections are sometimes indistinguishable from Phanerozoic glacial sections and were first erroneously referred to the Phanerozoic Eon (for instance, the Vendian glacial sequences of Middle and Central Asia were first considered as Carbonaceous–Permian in age).

The Late Vendian–Early Cambrian (Baikonur or West African) glaciation was complicated by climatic oscillations. Good examples are glacial sections in the Mauritanian Ardar composed of two main continental glacial formations separated by fluvial, lagoonal, and marine sediments and by unconformities (Deynoux and Trompette, 1981), which imply two glacial epochs at least. Even more intricate scenario is inferred for that glaciation from sections of western Mali by Proust and Deynoux (1994). They distinguished glacial epochs (a few million years long) and Milankovitch cycles approximately 100 ka long in the West African "glacial period" (20–30 m.y. long).

Similar climatic fluctuations are also recognizable in the Early Vendian Laplandian (Varangian) Glacial period. The Vil'chitsy Group of Belarus of that period consists of two units: the lower, Blon' Formation and unconformably overlying Glussk Formation (Fig. 4c). The first unit is composed of glacial sediments in its lower part and of interglacial sandstones and sandy dolomites in the upper part. In its most complete sections, the Glussk Formation encloses three till members separated by varved clays with dropstones and by well-sorted, probably fluvial sands with thin clay intercalations, rare ripple marks, and desiccation cracks (Fig. 4b). The uppermost layers of these members reveal glacial dislocations. Each member is of an intricate structure and consists of several layers different in composition and facies affinity, which are usually separated from each other by thin-bedded clay with glacial dislocations (Fig. 4a). It may be assumed that the Glussk and lower Blon' formations correspond to glacial epochs, whereas intervening strata characterize an interglacial epoch of the Laplandian Glacial period. In the considered case, combinations of different epochs represent long-period climatic fluctuations. A similar intricate three-unit structure in general is characteristic of the stratotype section of the Laplandian Glacial period in northern Norway, sections of Spitsbergen (Chumakov, 1978), eastern Greenland (Hambrey and Spencer, 1987), one of the most complete Laplandian sections in the Middle Urals (Chumakov, 1998), and Lower Vendian Nantuo glacial formation in southern China (Chumakov and Sergeev, 2004). Even more intricate are the Lower Vendian Fiq Formation in Arabia (Leather *et al.*, 2002), Blaini Formation in India (Kumar *et al.*, 2000), Yerelina Subgroup in Australia (Preiss, 2000), Puga and Bauxi subgroups in Brazil (Alvarenga and Trompette, 1988), which consist of

three and more subformations or till members separated by non-glacial sediments.

The alternating tillites and members of lacustrine and fluvial sediments within the Glussk Formation suggest three events of glacier advance and retreat. These events may correspond to glacial and non-glacial ages, and their combinations characterize medium climatic fluctuations. Similar members and climatic events can be outlined, with a variable confidence degree, in other abovementioned Laplandian sections as well.

The most complete successions of the first and second Late Riphean glacial periods are usually composed also of alternating glacial and interglacial sequences, which suggest different climatic epochs. For instance, sediments of the last Late Riphean glaciation in southern Australia (Sturtian after Preiss, 1987) and in the western United States (Pocatello and Perry Canyon groups and their analogues; Link *et al.*, 1994) show indications of two large glacial events, which can be considered as glacial epochs consisting of shorter glacial episodes. The same is typical of two glacial subunits of the Grand Conglomerate in Katanga (Hambrey *et al.*, 1981), which corresponds probably to the first Late Riphean glacial period.

The Early Proterozoic and Late Archean. Signs of large climatic fluctuations are distinguishable also in the Early Proterozoic. Sections of the Canadian Glacial period in North America are up to 8 km thick in total, composed of two lower groups and of the lower part of the third group of the Lower Proterozoic Huronian Supergroup (Fig. 5b). Three glacial formations, the basal units in these groups, are separated by interglacial sequences, which are 1.5 to 3.5 km thick, consisting each of two formations. As was noted, the Canadian Glacial period lasted approximately 100 m.y. Therefore, subordinate glacial events, which resulted in accumulation of the abovementioned glacial formations, should logically be considered as glacial epochs. Each of glacial formations is of an intricate structure: par example, the Gowganda Formation encloses three glacial members from 80 to 150 m thick, which are composed of several different tillite beds separated by unconformities, shales, and sandstones (Fig. 5a). Events corresponding to these members can be identified with analogues of glacial ages and their alternation with interglacial sediments as the middle-rank climatic fluctuations. Sufficiently intricate is the section of the Canadian glacial period in South Africa, where boreholes recovered six glacial members of the Griquatown glacial unit, which are from 8 to 95 m thick and separated by interglacial members of sandstones, ferruginous carbonates, and limestones from 4 to 16 m thick (Visser, 1981). Similar intricate structure is characteristic of this unit in the Transvaal basin as well (Reitfontein Diamictite Member).

The alternation of several glacial and interglacial members is also observed in Upper Archean glacial sections of the South African Republic. For instance,

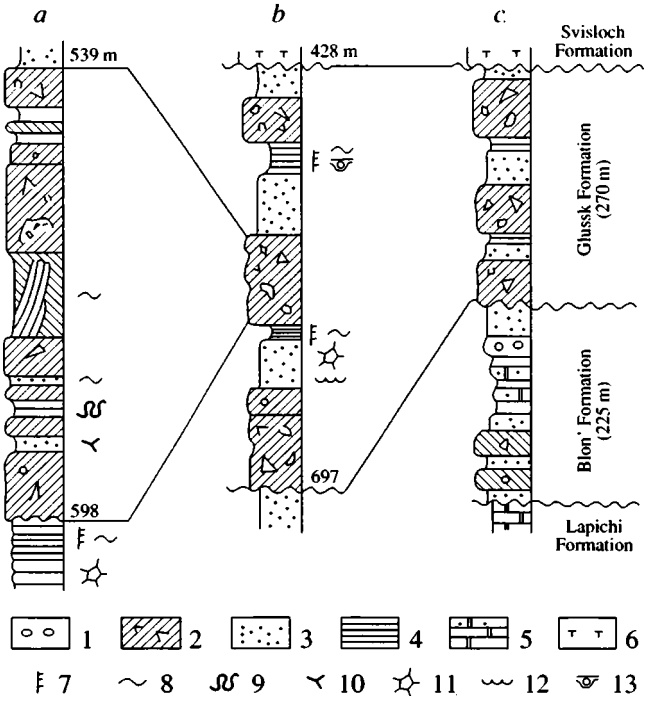

Fig. 4. Cyclic structure of the glacial Vil'chitsy Group, Lower Vendian, Belarus: (*a*) middle member of the Glussk Formation; (*b*) Glussk Formation; (*c*) Vil'chitsy Group; (1) conglomerate; (2) till; (3) sand; (4) clay and siltstone; (5) dolomite and sandy dolomite; (6) tuffite; (7) varved lamination; (8) glacial dislocation; (9) cryoturbation; (10) bioturbation (?); (11) desiccation cracks; (12) erosion signs; (13) dropstones.

the glacial part of the Mozaan Supergroup encloses four glacial beds from several to 20–30 m thick, which are separated by sandstone and shale members tens of meters thick (Young *et al.*, 1998).

Thus, it can be assumed, with a great confidence, that climatic fluctuations of highest ranks (superlong, long, and, probably, middle) were characteristic not only of the Phanerozoic, but of the Precambrian glacial periods as well. There are grounds to believe that there were also shorter climatic oscillations in the Precambrian.

3. GENERAL FEATURES OF CLIMATIC FLUCTUATIONS

The following peculiarities of climatic fluctuations can be distinguished within several large intervals of geological history.

Stability of climatic fluctuations hierarchy. The aforementioned data indicate that the multilevel hierarchy of climatic fluctuations maintained its stability during the entire Phanerozoic, i.e., over 500 m.y. Within glacial intervals of the Vendian and Late Riphean, recognizable are superlong, long, and middle fluctuations; short oscillations are also not inconceivable. Similar

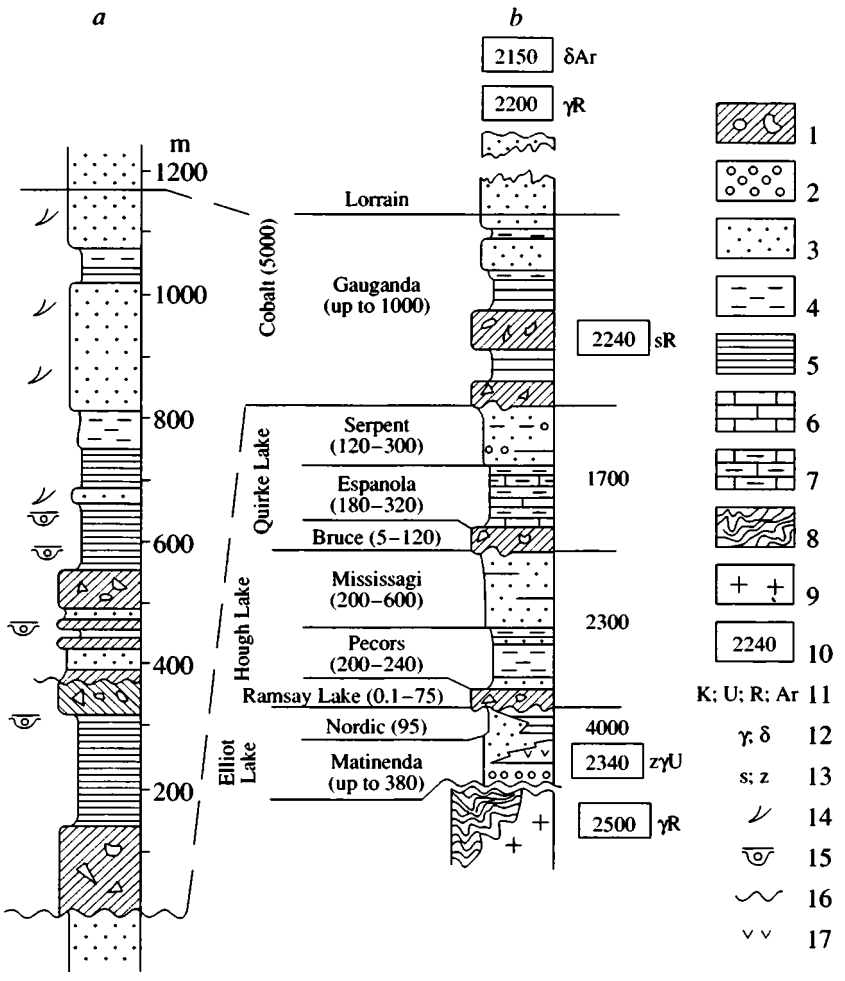

Fig. 5. Cyclic structure of the glacial Gauganda Formation (*a*) and Huron Supergroup (*b*), Lower Proterozoic, Canada (after Young, 1970): (1) tillite; (2) conglomerate; (3) sandstone; (4) siltstone; (5) argillite and shale; (6) limestone; (7) clayey limestone and marl; (8) crystalline rocks of the basement; (9) granite; (10) radioisotope age; (11) K–Ar (K), U–Pb (U), R–Sr (R), and Ar–Ar (Ar) radioisotope dating methods; (12) granite-based (γ) and diabase-based (δ) dating; (13) mica-based (s) and zircon-based (z) dating; (14) cross bedding; (15) dropstones; (16) erosional surface; (17) volcanics.

climatic fluctuations were characteristic as well of the Early Proterozoic and Late Archean glacial intervals, although is difficult to estimate their rank so far.

The existence of a stable multilevel hierarchy of climatic fluctuations throughout the Phanerozoic at least implies numerous and stable factors governing the above fluctuations. Accordingly, there is a principal opportunity to define individual oscillations in integral paleoclimatic curves and to interpret relevant processes using characteristic parameters of fluctuations (frequency, amplitude, configuration).

Amplitudes of climatic fluctuations. It is logical to consider the amplitudes of climatic fluctuations as a sum of superimposed elementary oscillations. For most small-scale climate changes, such an assumption is likely correct. A great deal in climatic fluctuations was however determined by feedbacks induced by climate changes in the biosphere. Negative feedbacks could weaken these changes and positive ones, to the contrary, strengthen them. Therefore, everything was complicated. For instance, short (Milankovitch-type) fluctuations in the Mesozoic resulted in relatively insignificant climate changes. To the contrary, similar fluctuations in the Pleistocene and Late Paleozoic could be accompanied by immense glaciations. This striking difference can be explained in the following way. Because of prolonged cooling periods related to superlong and long climatic fluctuations, the axis of short-period oscillations approached the temperature threshold, after which there were the onset and rapid formation of ice shields (Chumakov, 1995b). Thus, during short (Milankovitch-type) cooling periods, this threshold could be crossed that triggered formation of ice shields. The subsequent rapid growth of albedo and other processes connected by a strong positive feedback with glaciations (Fig. 6) stimulated further the temperature fall and substantial increase in short-period cooling amplitudes.

From the standpoint of above assumptions, the following working hypothesis can be proposed to explain different scales of climate changes. The low-amplitude fluctuations dominate during the non-glacial climate and represent a sum of changes produced elementary climatic oscillations at a given moment. In this case, the integral effect of changes is determined by cophasing of elementary climatic oscillations. When integral climate changes are significant, their final amplitudes can be strongly influenced via positive feedbacks by processes in biosphere. The formation of perennial glaciosphere was an extreme case of positive feedbacks. In such a case, amplitudes of climatic fluctuations increased stepwise.

Prevalence of non-glacial climate. Non-glacial (thermal) periods spanned approximately 70% of the Phanerozoic time, 78% of the last 1000 m.y. with periodic glaciations, and almost 90% of the preceding 2000 m.y. with episodic glaciations. The real role of non-glacial climate in geological history was even greater, because glacial episodes were discrete and alternated with interglacial episodes at all levels, from glacial periods to glaciations s.str. (or "glacial stages" in terminology of Quaternary geologists). Therefore, considering glacial and non-glacial episodes as equal in duration at first approximation, we can calculate that the integral duration of glacial epochs was approximately 50 to 60% of glacial periods. Only 50 to 60% of glacial epochs themselves were represented by glacial ages, which consisted, in turn, of glaciations s. str. and interglacial events. Judging from temperature curves compiled for the last 420 ka based on deuterium- and oxygen-isotope data for the Antarctic ice cores (Kotlyakov and Lorius, 2000), glacial conditions in high latitudes lasted approximately 85% of the second half of the Pleistocene glacial age. This value is likely the highest one characterizing the polar regions during the glacial maximum. Thus, it can be easily calculated that glaciations s. str. spanned not more than 30% of glacial periods and, consequently, not more than 10% of the Phanerozoic, 7% of the last 1000 m.y., and 3% of the preceding 2000 m.y. These values specified based on recent stratigraphic data define more exactly previous slightly higher estimates (Chumakov, 1995b).

The prevalence of non-glacial climate and subordinate role of glacial climate throughout the geological history suggest that global temperature on the Earth fluctuated mainly around positive values. In other words, the axial line of these fluctuations was within the interval of positive global temperatures. This can be illustrated by the following example, which allows simultaneously an approximate estimate of the global temperature on the glacial and ice-free Earth. We live in the glacial period. This is unambiguously indicated by the existence of significant polar ice caps and thick planetary psychrosphere. As is known, the average present-day temperature of the Earth surface is close to +15°C. During the last glacial maximum, this parameter was by several degrees lower and during the Late

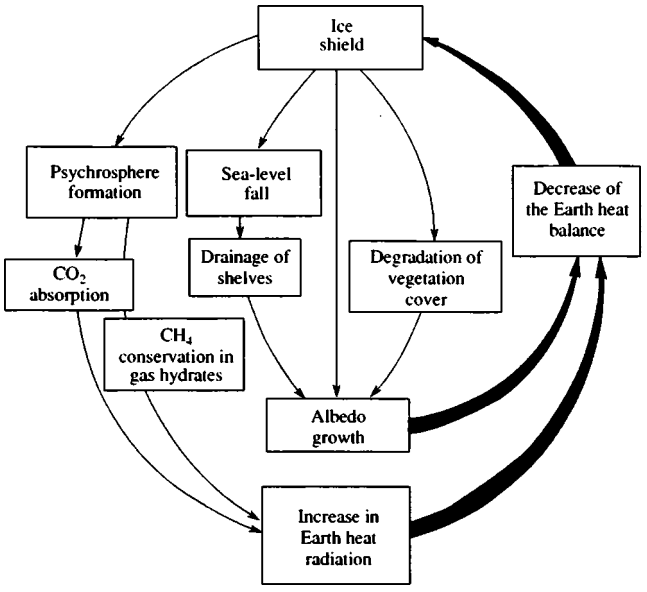

Fig. 6. Positive feedbacks in the ice shield–biosphere system.

Cretaceous non-glacial epoch, by several degrees higher as compared with the present-day value, i.e. in both cases it was well within the interval of positive temperatures. Mathematical modeling shows that the greatest part of the World Ocean was ice-free during the Late Precambrian glaciations and consequently, the average temperature of the Earth surface remained above zero, even though the land glaciers could form in low latitudes (Poulsen *et al.*, 2002; Poulsen, 2003).

The temperatures, around of which climatic fluctuations were scattered, changed with time. The main factors, which could change them, were planetary albedo, composition of the atmosphere, and orbital parameters. It is logical to assume that the temperature threshold, after which the perennial glaciosphere formed and non-glacial climate gave way to the glacial one, was not constant as well. It depended, to a certain extent, on paleogeographic situation in high and middle latitudes, as well as on circulation systems in the hydrosphere and atmosphere.

Global, synchronous, and cophasal climatic fluctuations. The approximate synchronism of glacial periods on different continents and, hence, their global scale is undoubted, confirmed by paleontological and radiometric dating. It is however more difficult to correlate climatic fluctuations of lower ranks in remote regions. This is possible only in case of stratigraphically well-studied intervals of the Phanerozoic. The correlation within these intervals shows that global and regional events are distinguishable among shorter climatic fluctuations.

Global fluctuations were synchronous and cophasal. This is distinctly exemplified by modern glaciers (Zakharov *et al.*, 1999; Solomina, 1999), by short and

ultrashort fluctuations during the historical period (Broecker and Denton, 1989), Holocene, and Pleistocene (Broecker and Denton, 1989; Thompson *et al.*, 1997; and others), by long-period fluctuations, and by general climatic trends during the Cenozoic (Fot'janova and Serova, 1994; Dingle and Lavelle, 1998), Cretaceous (Krasheninnnikov *et al.*, 1990; Herman, 1993; Frakes, 1999), and Permian (Chumakov, 1995a). It is noteworthy that cophasing of these fluctuations is traceable in various latitudes (low ones included) of both hemispheres. For instance, the medieval Little Glacial "Period" and preceding warm Vikings "Period" were manifested as respective cooling and warming near the North Polar Circle and in the Sargasso Sea (Keigwin, 1996). The Holocene ultrashort climatic fluctuations are registered in polar regions and Equatorial Africa (Stager and Mayewski, 1997). The last glacial maximum of the Pleistocene is recorded as cooling in low latitudes. In equatorial South America, this event is inferred from lowering of the snow line and feeding areas of mountainous glaciers by 700–1200 m (Broecker and Denton, 1989) and from oxygen-isotope characteristics of their ice. The cooling is evident from geochemistry of noble gases in underground water of equatorial Brazil and lowered snow lines and vegetative belts in Equatorial Africa, Sumatra, and New Guinea (Ninglan *et al.*, 1999). The penultimate interglacial and last glacial episodes are distinguishable in high and middle latitudes, and corresponding warming and cooling are established in Java (van der Kaars and Dam, 1995). During the Late Cretaceous and Cenozoic, the general cooling trend and long-period climatic fluctuations were similar, although differing in amplitude, in high and middle latitudes of Northeastern Asia (Krasheninnikov *et al.*, 1990; Herman, 1993; Velichko *et al.*, 1994; Fot'janova and Serova, 1994; Zakharov *et al.*, 1999), in surface and deep oceanic waters of low latitudes (Douglas and Woodruff, 1981), in middle latitudes of the southern Indian Ocean (Clarke and Jenkins, 1999), and in Antarctica (Dingle and Lavelle, 1998). Almost everywhere in these regions, warming is noted for the Cenomanian, Santonian, Campanian, early–late Eocene, and middle Miocene intervals, whereas cooling is registered in the Maastrichtian, terminal Eocene–Oligocene, and in the second half of the Miocene to Pliocene (Fig. 7).

The mentioned data imply that many climatic fluctuations (ultrashort, short, long, and superlong) were of global scale, synchronous and cophasal. This leads to the important conclusion that global climatic fluctuations represented a response to changes in the heat balance of the Earth (Chumakov, 1995a), rather than a result of changes in the mechanism of heat redistribution in the biosphere as its is commonly assumed (*CLIMAP...*, 1976; Nikolaev, 2000; and others). It should be emphasized that a thorough study of global fluctuations is very important for paleoclimatic interpretations. Their identification narrows the possible sphere of factors responsible for global climate changes.

Regional climatic fluctuations. Fluctuations of such kind are observable now and, undoubtedly, were characteristic of many regions in the past. As is known from meteorology, geography, and historical geology, they are induced by macrogeographic factors (changes in position, size, and configuration of continents, seas, and oceans, in the topography and landscapes), which redistribute heat in the biosphere. Past regional climatic fluctuations can be exemplified by the Turonian–Coniacian warming in Alaska that occurred in response to opening and widening of the fore-Cordillera strait ("Western Seaway") and by a relative cooling in Northeastern Asia caused by closure of a seaway between Asia and Alaska (Spicer and Herman, 1998). The other examples of regional climatic changes related to seaways opening and closing are climatic fluctuations in West Siberia and the Arctic region during the Paleogene and Neogene (Akhmetiev, 1996).

4. TRANSFORMATION OF CLIMATIC ZONALITY

4.1. Global, latitudinal, and regional climates. Global climatic fluctuations resulted in transformations of latitudinal zonality, which were different in scale and duration. Sufficiently well studied are changes in climatic zonality of the Quaternary (Zubakov, 1986; Barash *et al.*, 1989; Velichko and Nechaev, 1999; and others), Cretaceous (Ronov and Balukhovskiĭ, 1981; Krassilov, 1985; Naidin *et al.*, 1986; Vakhrameev, 1988; Chumakov *et al.*, 1995), Permian, and initial Triassic periods (Ziegler *et al.*, 1997; Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003), which are good examples of zonality transformations. The transformations were mostly rather gradual, of a relatively small-scale. They changed somewhat the width, latitudinal position and climatic parameters of some belts. On the other hand, particular transformations were rapid, qualitatively significant, and they can be termed as reorganizations of the climatic belts systems. Under their influence, new belts appeared, some old belts disappeared or became reduced, and main belts changed their latitudinal position and width. These parametric changes were of prime significance for climate, because the latitudinal position of belts determines not only the intensity of solar radiation and average temperatures, but also the seasonality (thermal and light), baric parameters, wind direction, precipitation, evaporation, cyclone trajectories, and circulation (Chumakov, 1995b). The reorganizations gave rise to new types of climatic zonality. Therefore, in addition to different climates in latitudinal climatic belts (*latitudinal climates*), one should distinguish the *global climate*, i.e., the type of planetary climatic zonality, first of all, by a set, width, latitudinal position, and contrasting parameters of climatic belts.

Latitudinal climatic belts were influenced by macrogeographic factors such as dimensions and position of continents, mountainous areas, seas, and seaways. These factors substantially modified latitudinal climate. Deforming and complicating climatic belts, they transformed latitudinal belts into sublatitudinal ones simultaneously creating a system of *regional, sectorial* (marine and continental to a variable extent), and *altitude* climate varieties within the belts.

4.2. Glacial and non-glacial global climate. Beginning from the Late Archean, two main types of the global climate alternated in geological history: glacial and non-glacial. Despite the prevalence of the non-glacial climate throughout the geological history (see section 3), numerous short glacial episodes were characteristic of long historical periods, particularly of the Riphean–Phanerozoic, and, therefore, two global climates repeatedly changed each other. This was caused mainly by changes in the heat balance of the Earth and led to formation and degradation of perennial glaciosphere (ice shields, perennial sea ice and permafrost) and psychrosphere. The alternating climates changed albedo and concentration of carbon dioxide in the atmosphere, sea-level and oceanic temperature fluctuations, proportion of land and sea areas, integral moisture transfer and latitudinal temperature gradient, circulation system in the atmosphere and hydrosphere, distribution of atmospheric precipitation and vegetative cover. Positive feedbacks intensified these processes (Fig. 6) and led to additional changes in the average Earth surface temperature and in biosphere as a whole, and eventually to principal transformations of climatic zonality (Chumakov, 1995b).

Glacial climate. The present-day climatic zonality of the Earth formed in the second half of the Cenozoic. This zonality and its varieties are sufficiently well known owing to recent geographic, biogeographic, and oceanographic studies, and to the thorough research of Late Cenozoic and Quaternary geology. Although the climatic zonality during the last glacial maximums and interglacials is sufficiently well studied, its development at the beginning of glacial periods is poorly known yet. An important comparative material is inferable from evolution of the glacial climatic zonality during the Permian (Ziegler *et al.*, 1997, 1998; Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003) and in the Late Ordovician (Barnes and Williams, 1991; Frakes *et al.*, 1992; Crowell, 1999). Data on Permian glaciations are of particular interest. In fact, only these data provide an insight into processes in the biosphere during the large-scale warming episodes on the Earth and by transition from the glacial to non-glacial climate.

The glacial climatic zonality is characterized by existence of ice shields and glacial belts. During glacial maximums ("great glaciations") of the Phanerozoic, glacial belts of high to middle latitudes advanced sometimes to the latitude of 30° as, for instance, in South

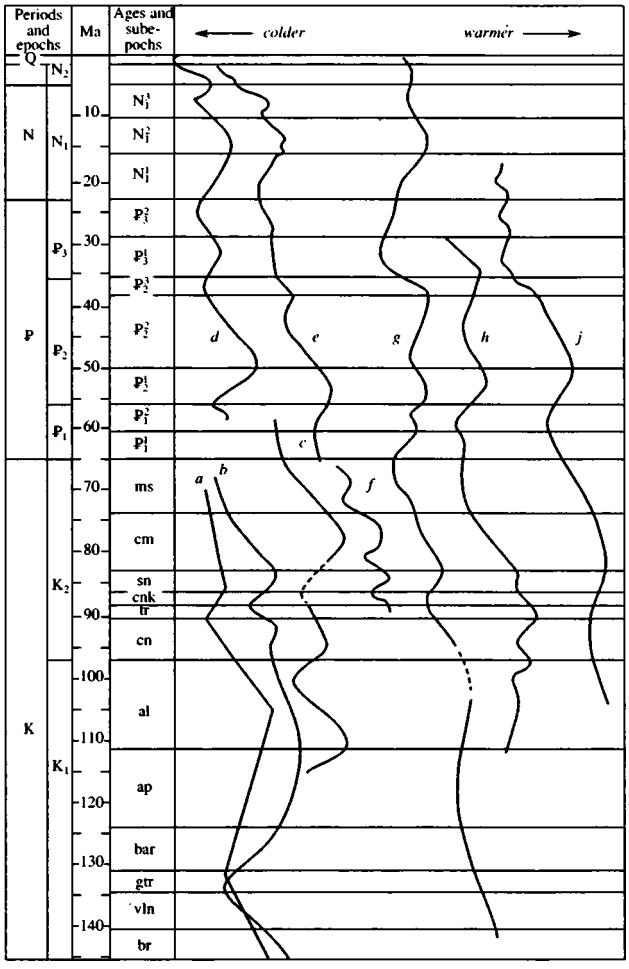

Fig. 7. Correlation of Cretaceous and Cenozoic climatic fluctuations in different latitudes of the Northern and Southern hemispheres: (a) conditional mean annual temperatures in Siberia inferred from oxygen-isotope and Ca/Mg data (Gol'bert, 1987); (b) relative temperature changes in northern Siberia inferred from lithological and paleontological data (Velichko *et al.*, 1994); (c) relative temperature changes in mountainous areas of northeastern Asia inferred from paleobotanic data (Krasheninnikov *et al.*, 1994); (d) relative variations of January temperatures in northeastern Siberia (Velichko *et al.*, 1994); (e) relative temperature changes in the Koryak Highland and Kamchatka inferred from paleobotanic and foraminiferal data (Fot'janova and Serova, 1994); (f) temperatures in the Northwest Pacific inferred from oxygen–isotope and paleobotanic data (Zakharov *et al.*, 1999); (g) surface water temperatures in low latitudes inferred from oxygen-isotope data (Douglas and Woodruff, 1981); (h) surface water temperatures in middle latitudes of the Indian Ocean inferred from oxygen-isotope data (Clarke and Jenkins, 1999); (j) relative temperature changes in the Antarctic Peninsula inferred from geochemical and other data (Dingle and Lavelle, 1998). Ages: (br) Berriasian, (vln) Valanginian, (gtr) Hauterivian, (bar) Barremian, (ap) Atpian, (al) Albian, (cn) Cenomanian, (tr) Turonian, (cnk) Coniacian, (sn) Santonian, (cm) Campanian, (ms) Maastrichtian.

America during the Asselian–early Sakmarian time of the Early Permian (Ziegler *et al.*, 1997; Chumakov and Zharkov, 2002, 2003; and Fig. 8a) or in the Late Ordovician. Temperate humid belts became reduced in these periods and even completely degraded sometimes, giving place to periglacial zones with cold steppe landscapes (tundra–steppes, after Philips, 1986; "periglacial steppes, after Velichko and Nechaev, 1999). In lower latitudes, the cold steppes gave way to the belt of semideserts and deserts with temperate to tropical climate and, sometimes, to cold semi-deserts as, for example, in South America during the Early Permian (Fig. 8a). As is shown in available reconstructions for the last Pleistocene glacial maximum, a strong reduction and transformation involved also subtropical belts, which were almost completely replaced by steppe, semi-desert, and desert landscapes with a temperate climate. Although climatic belts of low latitudes escaped such a drastic transformation, zones of deserts and semi-deserts substantially widened there at the expense of subequatorial belts (savanna), whereas savannas replaced partly the former humid equatorial belts (Phillips, 1986). Widening of arid and semiarid belts at the expense of neighboring belts is also established for the Early Permian glacial maximum. The global climate type under consideration can be termed as the *climate of glacial maximums* or *climate of great glaciations.*

The present-day climatic zonality allows us to imagine the past global climate of interglacial epochs. Data on the Early Permian climate confirm such a possibility. In addition, they suggest that initial and terminal stages of glacial periods were also characterized by the climatic zonality similar to the present-day one. Retreated from middle latitudes during the Late Sakmarian time and contracted to the size of the present-day polar cap, the tremendous Gondwanan ice shield still existed, although it was oscillating, in the Artinskian Age and probably later (Zharkov and Chumakov, 2001b; Chumakov and Zharkov, 2002). The Permian polar cap resembled the present-day one being similarly surrounded by periglacial zones and temperate cold forests, which were comparable with taiga in opinion of some experts in paleobotany. The glacial belt reduction was accompanied by reviving of the temperate cold humid belt in the Southern Hemisphere and by a slight widening of similar belt in the Northern Hemisphere. An insignificant reduction was characteristic of the northern arid belt, while its southern counterpart widened. Thus, the Permian epoch was characterized by a substantial reorganization of the climatic zonality as compared with that of the glacial maximum. The global glacial climate similar to the present-day or Late Sakmarian–Early Artinskian climate can be termed as the *climate of polar caps.* It was likely characteristic of the other Pleistocene interglacials and, probably, of the Late Eocene, Oligocene, and Neogene minimums, when the Antarctic glacial cap existed.

Paleoclimatic reconstructions for the Kazanian–Tatarian ages of the Late Permian illustrate the climatic zonality during the terminal phase of the Carboniferous–Permian glacial period. The southern polar cap degraded by that time, although some active glaciation centers that produced icebergs were preserved in Antarctica and southeastern Australia. In northeastern Asia, glaciation centers appeared again indicating continuation of the glacial period on the Earth. Nevertheless, the former climatic zonality changed, as a continuous glacial belt disappeared and polar areas became moderately cold in general. They were rimmed by belts with a temperate humid climate. The climatic asymmetry of hemispheres substantially decreased and synonymous climatic belts were more symmetrical relative to the equator (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003). This variety of the glacial climate can be termed as the *climate of cold polar areas.* The next climatic reorganization induced by a sharp global warming near the Permian–Triassic boundary resulted in replacement of the glacial climate by the non-glacial one.

Non-glacial climate. In distinction from the glacial climate, this type of the global climate is distinguishable based on the past geological examples, i.e. on paleoclimatic reconstructions only, because of impossibility to apply actualistic climatic models. The reconstructions indicate that the non-glacial global climate was also non-uniform. At present, it can be subdivided in two, the *non-glacial humid* and *non-glacial arid* varieties (Zharkov and Chumakov, 2001a).

The typical non-glacial humid climate prevailed in the Late Cretaceous beginning from the terminal Albian. The climatic zonality peculiar of this climate is considered in several works (Ronov and Balukhovskiĭ, 1981; Vakhrameev, 1988; Krassilov, 1985; Chumakov *et al.*, 1995; Zharkov and Chumakov, 2001a; Valdes *et al.*, 1999; Barrera and Johnson, 1999). Polar areas in both hemispheres were occupied at that time by temperate and temperate warm belts with positive annual temperatures (Ditchfield *et al.*, 1994; Spicer *et al.*, 1996, Kennedy, 1996) and very rare frosts (Falcon-Long *et al.*, 2001). Widespread in these belts were deciduous broad-leaved and coniferous forests, where dinosaurs herded, and lakes populated by crocodiles (Vakhrameev, 1988; Chumakov *et al.*, 1995; Clemens and Nelms, 1993; Tarduno *et al.*, 1998; Falcon-Long *et al.*, 2001). The warm humid belts were substantially wide, occupying the middle (Fig. 8c) and partly the high latitudes of both hemispheres (Cenomanian and Turonian ages). Judging from occurrence of the monoxylic wood remains, the climate in these belts was mostly frost-free. The arid belts located in tropical and, partly, middle latitudes were moderately developed. At the beginning of the Late Cretaceous, they occupied approximately 35% of land, and their area gradually reduced to 25% by the end of this epoch, (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2003; Fig. 8b) giving way to the widening humid equatorial belt. In total, humid belts covered 75% of land at the end of the Cretaceous.

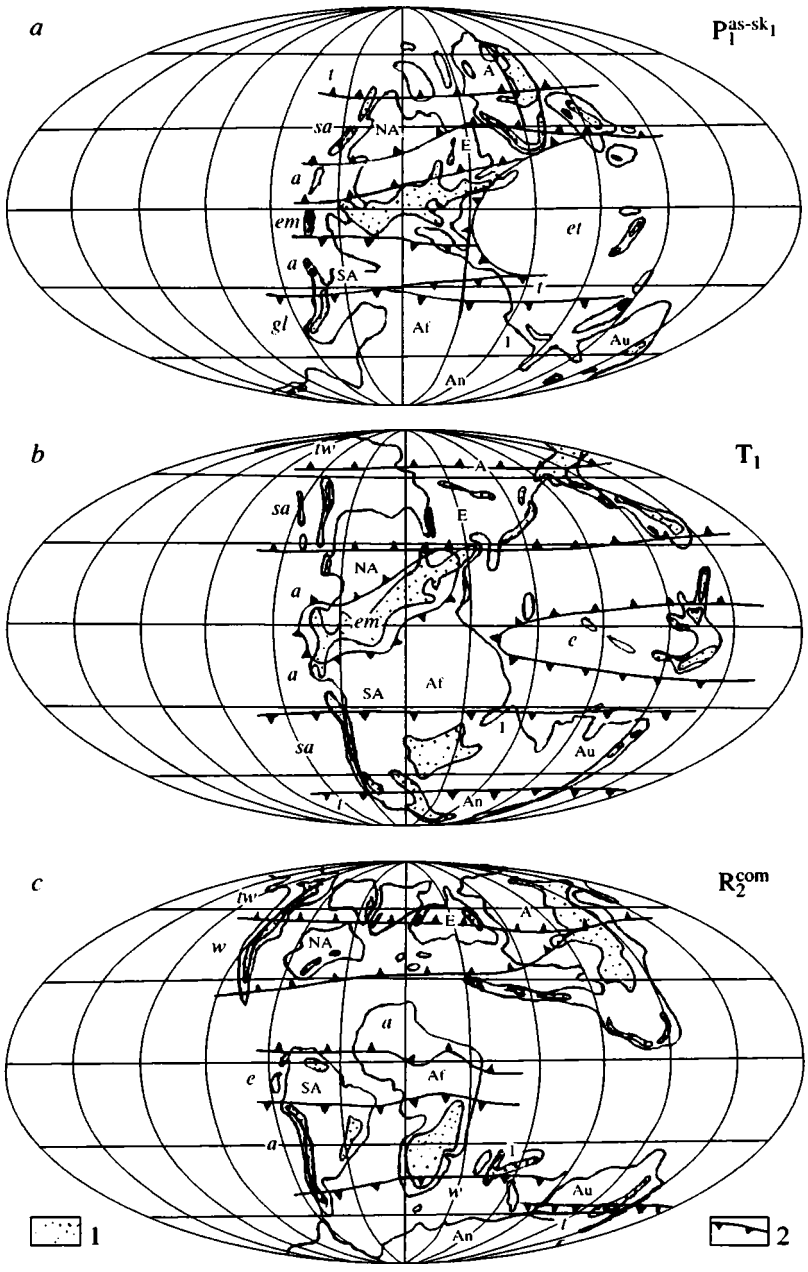

Fig. 8. Climatic zonality (a) during the Asselian–early Sakmarian time (after Chumakov and Zharkov, 2002), (b) in the Induan Age of the Early Triassic (after Chumakov and Zharkov, 2003), (c) in the Maastrichtian Age of the Late Cretaceous (after Zharkov and Chumakov, 2001a): (1) mountainous belts and regions; (2) boundaries of climatic belts (after Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003). Continents: (A) Asia, (Au) Australia, (An) Antarctica, (Af) Africa, (E) Europe, (I) India, (NA) North America, (SA) South America. Climatic belts: (*gl*) glacial, (*t*) temperate, (*tw*) temperate warm, (*w*) warm, (*sa*) semiarid, (*a*) arid, (*et*) equatorial–tropical, (*e*) equatorial, (*em*) equatorial mountainous.

The Induan Age of the Triassic was a time of typical non-glacial arid climate (Zharkov and Chumakov, 2001a, 2001b; Chumakov and Zharkov, 2003; Fig. 8b). The humid temperate belts of northern and southern high latitudes and the equatorial humid belt (mountainous areas with a vertical climatic zonality) occupied, in total, 20% of land at that time. The remainder was covered by arid and semiarid belts covering almost all the middle and partly high latitudes (Fig. 8b). Semiarid belts were particularly widespread. They were about 40° wide in both hemispheres. The narrow equatorial belt was discontinuous and, therefore, the northern and southern arid belts joined locally each other. The typical equatorial humid climate was preserved at that time only within the Cimmerian arc that represented an archipelago of microcontinents between the Paleo- and Neo-Tethys.

The genesis of the Induan Age climate represents a certain problem. An increase of the global surface temperature should be theoretically favorable for elevated evaporation in oceans and seas and, correspondingly, for a higher humidity of the atmosphere (Barron *et al.*, 1989). This is likely correct for the Late Cretaceous epoch. Nevertheless, the situation was more ambiguous, because the moisture distribution over land was largely controlled by macrogeographic peculiarities of the Earth.[4] A combination of these factors determined the remoteness and latitudinal position of moisture sources, paths of moisture transfer in the atmosphere, and distribution of baric and orographic barriers along them. Therefore, the breakup of supercontinents and opening of new oceans coupled with transgressions and wide distribution of epicontinental seas and seaways favored development of the non-glacial humid climate in the Late Cretaceous, while a combination of mountainous supercontinents with regressions in marginal orogenic belts resulted in prevalence of non-glacial arid climate at the beginning of the Triassic (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003).

4.3. Asymmetry of climatic zonality. Typical of the present-day climatic zonality of the Earth is a slight asymmetric position of climatic belts relative to the equator. Similar asymmetry is likely a usual phenomenon characteristic of planets with the atmosphere. It is characteristic of Mars, par example, the polar caps of which are substantially different in size. The climatic asymmetry of the Earth is thought to be a consequence of asymmetrical position of continents, oceans, seas, and, correspondingly, of circulation systems in the Northern and Southern hemispheres. Moreover, these systems are autonomous to a certain extent. Inasmuch as land, oceans, and seas were hardly located symmetrically relative to the equator at any time before, one can assume that the climatic asymmetry of hemispheres existed throughout the geological history. Paleoclimatic reconstructions confirm this assumption. An insignificant climatic asymmetry existed in various forms during the entire Cretaceous period. At that time, the warm humid belt in the Southern Hemisphere (dissimilar to the Northern one) reached sometimes (Neocomian, Cenomanian, Turonian) the high latitudes (Vakhrameev, 1988; Chumakov *et al.*, 1995). The arid belt of the Southern Hemisphere was slightly wider in the Early Cretaceous, Cenomanian, and Turonian, whereas the equatorial humid belt advanced noticeably southward in the Maastrichtian. It seems that, dissimilar to its present-day state, the Southern Hemisphere was slightly warmer in the Cretaceous Period than the Northern Hemisphere. At the beginning of the Triassic, climatic asymmetry was weak, like in the Cretaceous and nowadays, but the Southern Hemisphere was slightly colder (Chumakov and Zharkov, 2002, 2003).

The present-day zonality and examples mentioned above imply that climatic asymmetry in periods of the interglacial and, particularly, non-glacial climate was usually insignificant even in case of highly asymmetrical position of land and oceans as it was characteristic, for instance, of the Early Triassic.

In periods of the glacial climate, particularly during "great glaciations," asymmetry in position of climatic belts substantially increased and was multiply manifested in its extreme form of the "monopolar glaciations". Although immediate causes of climatic asymmetry on the Earth might differ, asymmetrical position of continents, oceans, large orogenic belts, and oceanic currents were the main factors responsible for this phenomenon. In the Pleistocene, ice sheets advanced to 38°N in the Northern Hemisphere and only to 60°S in the Southern Hemisphere, because they were kept back by oceans surrounding Antarctica. The Late Ordovician glaciations developed only the Southern Hemisphere, because there was no land in high and middle latitudes of the Northern Hemisphere. The most remarkable example of extreme climatic asymmetry and its dependence on glaciation was a climatic zonality of the Asselian–Early Sakmarian time (initial Early Permian) (Chumakov and Zharkov, 2002). As was mentioned, the glacial belt occupied almost entire high and middle latitudes of the Southern Hemisphere at that time. The southern temperate humid belt was reduced to a narrow (10°) westward pinching-out band, and the semiarid belt disappeared (Fig. 8a). To the contrary, the Northern Hemisphere lacked the glacial belt. The temperate belt of this hemisphere occupied the high and nearly all the middle latitudes, while semiarid belt was as wide as 20°. Despite a similar disposition of land, seas, and oceans at the beginning of the Triassic, when the glacial climate on the Earth was replaced by the non-glacial one, climatic asymmetry immediately became minimal (Fig. 8b).

CONCLUSION

The discussed trends in climate changes are mainly inferred from geological data being consequently empirical. At the scale of an eon, they outline irreversible changes from non-glacial climate of the Early Archean to quasi-periodic glacial climate of the Late Riphean, Vendian and Phanerozoic. The stable hierarchy of global, synchronous, and cophasal climatic fluctuations of 10–12 ranks is distinguishable within the general gradual climatic trend. The fluctuations determined gradual transformations or rapid reorganizations of the climatic zonality on the Earth depending on the scale of fluctuations and induced feedbacks.

This implies that evolution of the Earth climate was relatively regular, not chaotic. The inferred regular climatic changes seem unexpected for such an intricate

[4] In addition to main planetary trends in moisture distribution, which are controlled by the form and rotation of the Earth and by circulation cells responsible for creation of primary latitudinal belts, the arid and humid ones inclusive.

open system as the climatic system of the Earth. In this connection, two assumptions are logical. First, the Earth climate system was quasistationary and this phenomenon was most clearly manifested during the large time intervals (eons, erathems, and periods) and by highest-rank fluctuations. Ultrashort and weather oscillations in particular were more chaotic, and this is not surprising. The non-linear development of natural processes, the geological ones included, usually changes depending on the analysis time scale (Grachev, 1998). Second, among all the diverse nonlinear processes that influenced the Earth climate there was likely a limited number of most influential "governing" processes with fixed or periodical attractors. These processes significantly differed from each other in the frequency of characteristic variations. The following "governing" processes can be outlined at present.

(1) Variations in the endogenic activity of the Earth were responsible for superlong (150 m.y.) and long (several tens of million years) climatic fluctuations. Contrary to the widely accepted standpoint, periods of superlong fluctuations substantially differ from duration of the galactic year. They were probably related to the self-oscillating geodynamic processes in the Earth interiors (Dubrovskiĭ, 1998; Dobretsov, 1999), although the influence of astronomical factors cannot completely be ruled out.

(2) The other factor is the carbonate and organic carbon burial in the sedimentary shell, lithosphere, and mantle of the Earth, the intensity of which depends, in particular, on geodynamics, latitudinal position of continents, weathering processes on land, and accumulation rates of organic carbon in seas.

(3) Variations in orbital parameters and inclination of the Earth axis are responsible for the Milankovitch-type short fluctuations. Their astronomical origin is inferred in many works and, what is important, this is supported by the fact that such fluctuations occurred throughout the entire Phanerozoic regardless of repeated changes in geological, geographic, and biotic situations on the Earth.

(4) Variations in the solar activity control the ultrashort climatic fluctuations.

There are probably the other "governing" processes, for example, those, which caused middle climatic fluctuations.

The other accompanying processes only modified, to some extent, the superlong and long climatic fluctuations determined by "governing" factors. They strengthened or weakened influence of the latter, being unable to control the alternation of glacial and non-glacial periods.

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Fusulinids and Permian Scale of the Tethys

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Abstract—The curve depicting generic diversity of Permian fusulinids has two distinct peaks: the Asselian–Sakmarian and Midian. The in-between minimum corresponds to the end of the Bolorian Age. The most significant extinctions of fusulinids are recorded in the terminal Midian (71% of all genera), Bolorian (48%), Asselian (27%), and Sakmarian (23%) ages. A renewal of prime importance took place in the Kubergandian Age, when 52% of new genera appeared in fusulinid community. These data suggest that history of Permian fusulinids can be divided in two principal (Asselian–Bolorian and Kubergandian–Dorashamian) and four second-order periods of evolution (Asselian–Sakmarian, Yakhtashian–Bolorian, Kubergandian–Midian, and Dzhulfian–Dorashamian). Two main periods of evolution corresponding to the Lower and Upper Permian of traditional East European scale are suggested to be termed as the Cisuralian and Tethysian subsystems. The subsystems are divisible into the Uralian, Darvasian, Yangsingian, and Lopingian series corresponding to second-order periods of fusulinid evolution. The suggested scheme is consistent with traditional two-part subdivision of the Permian System. In distinction from the three-series global chronostratigraphic scale accepted recently, series in the suggested scheme are more proportional and better corresponding to the natural stages in evolution of marine biota. The global scale with its strict criteria used to define boundaries between series and stages is hardly applicable in the Tethyan region. This is evident from diverse opinions concerning position of the Guadalupian Series lower boundary in Tethyan sections.

Key words: fusulinids, diversity, evolutionary stages, Tethyan scale, Permian System, subsystems, series, stages.

INTRODUCTION

After long-lasted discussions, the new global chronostratigraphic scale of the Permian System was officially authorized at the beginning of 2001. The system was divided into three series (Cisuralian, Guadalupian, and Lopingian) and nine stages. The Asselian, Sakmarian, Artinskian, and Kungurian stages are included into the Cisuralian Series; the Roadian, Wordian and Capitanian stages are regarded as subdivisions of the Guadalupian Series, and the Lopingian Series is divided into Wuchiapingian and Changhsingian stages. The scale is primarily based on distribution of conodonts, which are rare and inadequately studied in Tethyan areas, where the scale can be hardly applied therefore. This problem has been considered in detail in earlier work (Leven, 2001). The other important problem concerning the new scale is an obvious disproportion between three series. The lower Cisuralian Series (23.5 m.y. long) is identical to the Lower Permian of the classical East European scale, the middle Guadalupian Series (12 m.y.) includes the Ufimian and Kazanian stages coupled with the greater lower part of the Tatarian Stage, while the third Lopingian Series (9.5 m.y.) corresponds to the rest of the latter (Menning, 2001). With such a disproportion, it is impossible to operate with the Guadalupian and Lopingian series in the Boreal and Notal regions.

At the congress "Permian System of the Earth" of 1991, I proposed a compromise stratigraphic scheme (Leven, 1992), in which two discriminated subsystems correspond to the Lower and Upper Permian of the classical scale and are divided in two series each. Series of the upper subsystem have been regarded therewith as corresponding to the Guadalupian and Lopingian series of the new scale, while the lower subsystem was divided into lower and upper series spanning the Asselian–Sakmarian and the Artinskian (Yakhtashian)–Kungurian (Bolorian) intervals, respectively. A similar four-member subdivision of the Permian has been suggested at the same congress by Ganelin and Kotlyar who based their conclusions on distribution of brachiopods in sections of the Boreal areas. There was no opposition to these proposals, but they have not been taken into account during the subsequent work on the new scale, because subdivision of the Lower Permian in two individual series has not been practically substantiated.

Obvious disadvantages of the new global scale are convincing that we need in a special scale for the Tethyan region, where application of the former is inconvenient. Such a scale tested in many sections of various Tethyan areas is already known (Leven, 1980a, 1981). This work is aimed to substantiate additionally the stages of that scale and their boundaries taking into consideration new data, which have been accumulated

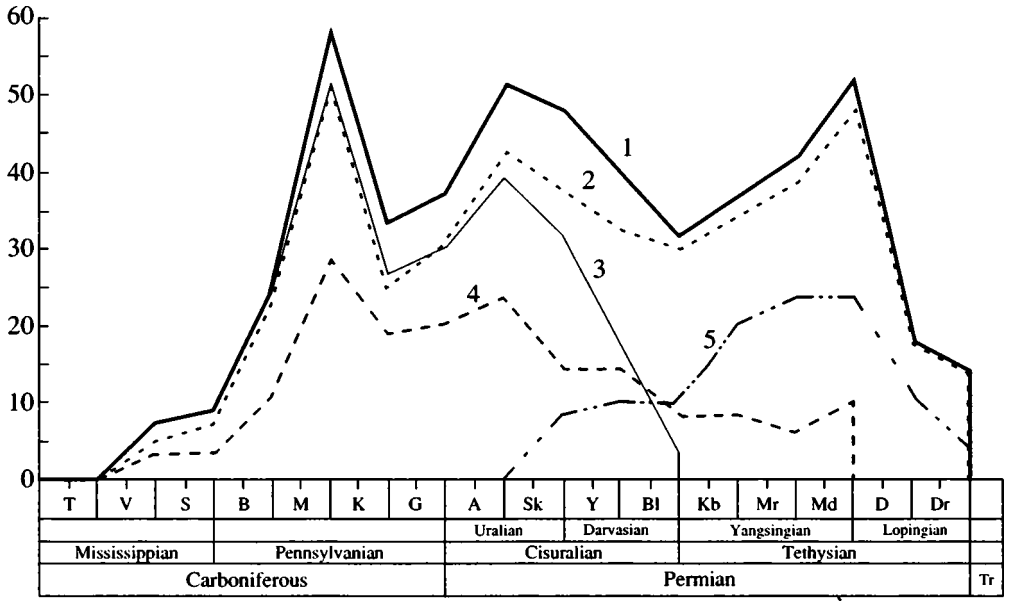

Fig. 1. Abundance rates of fusulinid genera calculated for the stage terminations: (1) quantity of genera worldwide; (2) Tethyan areas; (3) East Europe; (4) United States, Texas; (5) peri-Gondwanan part of the Tethys. Stage indices: (T) Tournaisian; (V) Visean; (S) Serpukhovian; (B) Bashkirian; (M) Moscovian; (K) Kasimovian; (G) Gzhelian; (A) Asselian; (Sk) Sakmarian; (Y) Yakhtashian; (Bl) Bolorian; (Kb) Kubergandian; (Mr) Murgabian; (Md) Midian; (D) Dzhulfian; (Dr) Dorashamian.

during the last two decades after the scale approval by the ISC. Simultaneously, the stage scale is coordinated with the suggested subdivision of the Permian System into subsystems and series.

DIVERSITY DYNAMICS OF FUSULINID GENERA

Fusulinids are most abundant, widespread and well-studied fossils characteristic of Permian deposits in Tethyan areas. Being of a high stratigraphic potential, they are used to elaborate the Permian stratigraphic scales, including the version considered below. Evolution of fusulinids was irregular, divisible into several stages of different orders. This is advantageous for hierarchic stratigraphic subdivision of the Permian System, beginning from zones to subsystems. Stages in evolution are recognizable based on various criteria, such as extinctions of different-rank taxa, diversity fluctuations, and successions of dominant fossil groups.

The presented diagrams illustrate the diversity variations of fusulinid genera beginning from their appearance in the Visean Age till the extinction at the end of the Dorashamian Age (Figs. 1, 2). The considered genera are classified in accord with "Guidebook to Systematics of Paleozoic Foraminifers" (Rauser-Chernousova *et al.*, 1996). The abundance rates of Tethyan fusulinids are calculated for either the undivided stages or the lower and upper substages separately. The curves depicting the secular diversity variations of fusulinid genera in Tethyan areas or worldwide (Fig. 1) are very similar, and this implies that Tethyan fusulinids define their general diversity.

In both curves, there are three diversity peaks, which took place in the Moscovian, Asselian–Sakmarian, and Midian periods. Accordingly, the entire history of fusulinid development can be divided in the Visean–Kasimovian, Gzhelian–Bolorian, and Kubergandian–Dorashamian periods of prime significance. The order Fusulinida was dominant during the first period. During the second one dominated fusulinids of the order Schwagerinida, while orders Neoschwagerinida, Staffellida and Schubertellida were the main taxa of the third period. The same conclusion is inferable from the histogram (Fig. 2) plotted based on calculations for the substage time spans. The only distinction is the two-peak character of the Asselian–Sakmarian maximum. The histogram also illustrates the abundance rates (in absolute values and percentages) of new and extinct genera for the each time span. Besides the Moscovian period of diversification, which is out the scope of this work, the appearance peaks of new genera are confined to the first half of the Kubergandian Age (52% of all genera) and to the second half of the Gzhelian Age (47%), i.e., to the commencement of second and third main periods of fusulinid development. Considerable renewals are also detectable in the early Yakhtashian (36% of new genera), Midian, and Asselian fusulinid assemblages. In addition to the complete extinction of fusulinids at the end of the Permian, catastrophic extinctions were characteristic of the terminal Midian (71% of all genera) and terminal Bolorian time (48%). Less significant extinctions were those signifying ter-

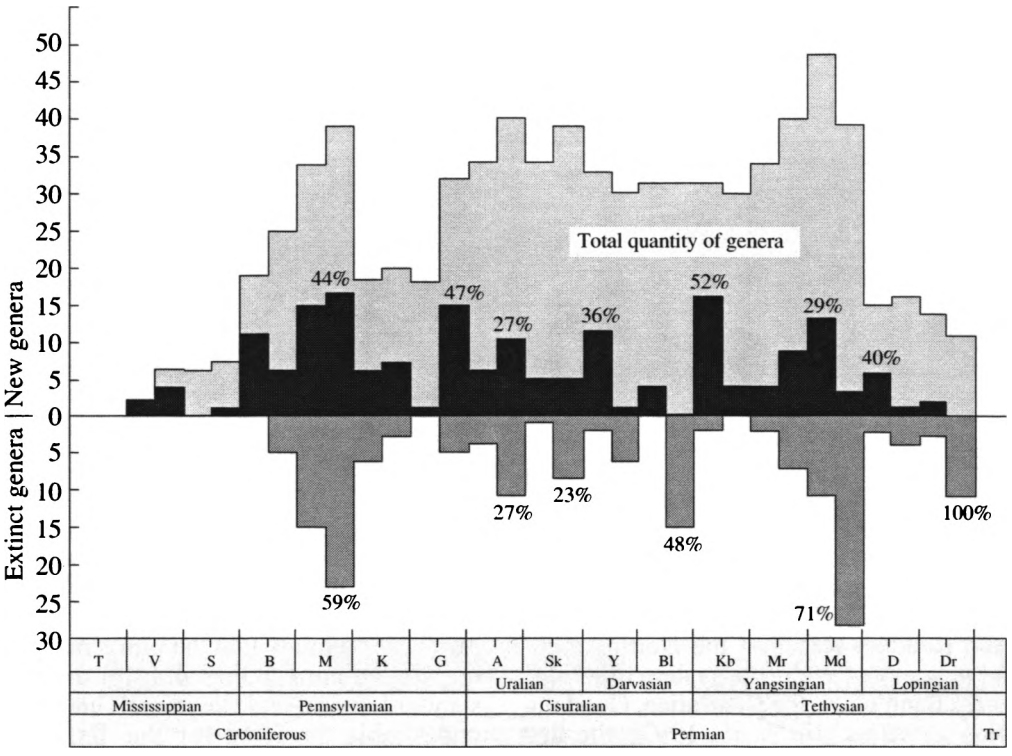

Fig. 2. Total quantity (shaded light gray) of fusulinid genera calculated per each half of an age; percentages of new (shaded black) and extinct (shaded dark gray) genera are calculated relative to the total quantity (stage indices as in Fig. 1).

minations of the Asselian and Sakmarian ages (27 and 23%, respectively). Consequently, the most abrupt and essential changes in Permian fusulinid assemblages are recorded across the Bolorian–Kubergandian and Midian–Dzhulfian boundaries.

The above data on distribution of fusulinids offer a possibility to substantiate subdivision of the Permian System into series and stages, at it is shown below.

BOUNDARIES OF THE PERMIAN SYSTEM

At present, the system lower boundary is placed officially at the base of the Asselian Stage in accord with the original idea of Ruzhentsev (1951) who recognized significance of this level based on distribution of ammonoids. In terms of fusulinid succession, this boundary marks the first appearance of the genus *Sphaeroschwagerina*. The genus that is widespread enables the boundary recognition in a vast territory from South and East European areas to Japan (Leven and Shcherbovich, 1978; *Guidebook...*, 1990; Davydov, 1995; Krainer and Davydov, 1998). In general, the fusulinid biota changed insignificantly across this boundary however. The more considerable changes are recorded in Gzhelian deposits. The changes were connected with diversification of the order Schwagerinida that appeared in the Kasimovian Age (47% of all genera) and gave rise to many new genera, the typical Lower Permian *Pseudofusulina, Rugosofusulina*, and *Dutkevitchia* included. Hence, it would be more reasonable to consider exactly this event as marking the Carboniferous–Permian boundary and to place this boundary somewhere in the Gzhelian Stage in its current understanding, or at the level separating the Gzhelian and Orenburgian stages, if the latter will be authorized. Such a viewpoint based on fusulinids complicates, however, the precise positioning of the boundary based on other fossils, e.g., on conodonts and ammonoids. The officially accepted boundary is more appropriate therefore.

According to decision of the International Stratigraphic Commission, the upper boundary of the Permian System is at the base of the *Hindeodus parvus* Zone of the conodont scale. The base of Bed 27c in the Meishan-D section (China) is selected for the GSSP of the boundary (*Vote on Meishan...*, 1999). Fusulinids disappear below this level and nowhere in the world attain it. In the indicated section, the last of them occur approximately 1 m below the Bed 27 (Zhao *et al.*, 1981). Thus, fusulinids do not fix precisely the upper Permian boundary that is placed slightly higher than the level of their final extinction.

SERIES OF THE PERMIAN SYSTEM

There are many variants of series discrimination in the Permian System. In the traditional East European scale, the system is divided in two main parts at the

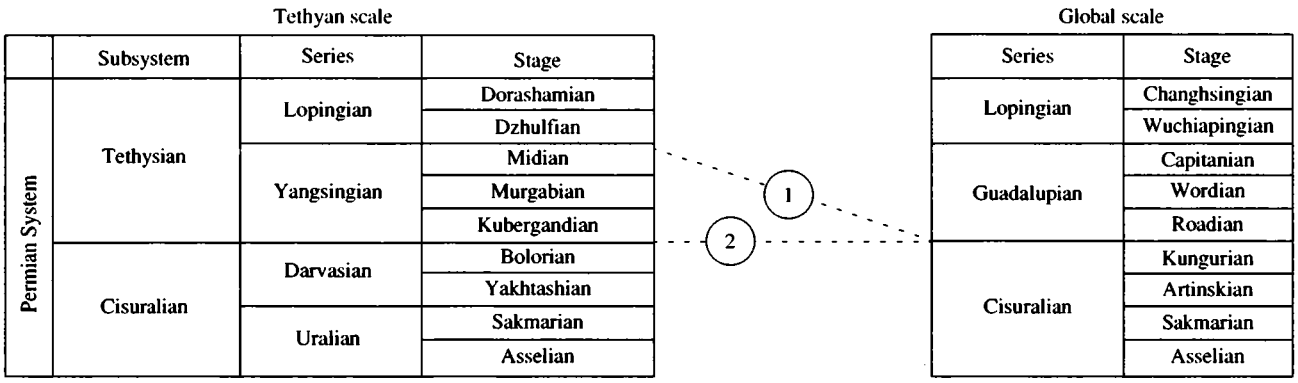

Fig. 3. Permian stratigraphic scale for Tethyan areas and its correlation with the global chronostratigraphic scale after Henderson *et al.*, 1999 (1), and after Leven, 2001, and Kozur, 1998 (2).

level corresponding to the boundary between Kungurian and Ufimian stages. In current nomenclature, approximately the same level separates the Bolorian and Kubergandian stages of the Tethyan scale or the Cathedralian and Roadian stages of the North American scale. In the new global scale, the system is divided however into three main units: the Cisuralian, Guadalupian, and Lopingian series (Jin *et al.*, 1997), the first and two second corresponding respectively to Lower and Upper Permian of the traditional East European scale.

From the viewpoint of main evolutionary periods of fusulinids, it is most reasonable to divide the Permian System in two parts corresponding to the second and third periods (Fig. 1), one characterizing the flourishing time of the order Schwagerinida, and the other one the prosperity time of orders Neoschwagerinida, Schubertellida, and Staffellida. Each of these periods can be divided then in two units of lower ranks spanning the Asselian–Sakmarian and Yakhtashian–Bolorian intervals of the Early Permian time and the Kubergandian–Midian and Dzhulfian–Dorashamian intervals of the Late Permian. In the global chronostratigraphic scale, the lower Cisuralian Series includes two former subdivisions, whereas the Guadalupian and Lopingian series correspond each to one of the latter subdivisions. The disproportion is obvious, and one can avoid it dividing the Cisuralian period in two individual series. As a result, the Permian System would be divided in two subsystems with two series in each (Fig. 3).

The name "Cisuralian" accepted in the new global scale can be given to the lower subsystem, and the upper subdivision of the same rank can be named the "Tethysian" Subsystem, because the most complete marine successions of this unit are known in Tethyan areas. Names "Uralian" and "Darvasian" appear to be appropriate for lower and upper series of the Cisuralian Subsystem, since they have been formerly used for stratigraphic subdivisions of approximately the same ranges (de Lapparent, 1900; Schuchert, 1935; Stepanov, 1951; Miklukho-Maclay, 1958). The upper subdivision of the Tethysian Subsystem is identical to the Lopingian Series of the global scale. As for the lower one, it could be named the Guadalupian Series, but the conodont dates (Mei *et al.*, 1998; Henderson *et al.*, 1999) imply that the unit's range, as I understand it (Fig. 3), corresponds only to the upper part of the standard series, and the name Guadalupian appears to undesirable therefore for the Tethyan successions. I propose to term it as the Yangsingian Series, because stratigraphic subdivision of nearly the same range is of a long-term usage under this name in China (Huang, 1932; Sheng and Jin, 1994).

Considered below are fusulinid assemblages characterizing four series outlined above, which correspond to second-order stages in evolution of the fossils under consideration (Plates I and II).

Cisuralian Subsystem

The Uralian Series spans the Asselian and Sakmarian stages of fusulinid peak diversities. The name "Uralian" was introduced for the so-called *"Schwagerina"* Horizon of the Ural region by de Lapparent (1900), who erroneously considered the horizon as a marine equivalent of the Stephanian Stage in the Upper Carboniferous succession of Western Europe. Ruzhentsev (1936) defined the *"Schwagerina"* Horizon as the Sakmarian Stage. He regarded the name "Uralian" as inappropriate for this stage, because different researchers used to be applying it for deposits of variable stratigraphic range. Following his argument, however, one should reject all the names of Permian stages, many of which are diversely understood even at present. The Sakmarian Stage distinguished by Ruzhentsev is not an exception, since its original range became reduced twice, when the lower half of the stage was attributed to the individual Asselian Stage.

The name "Uralian," approximately in its initial meaning, is suggested here for the lower series of the Cisuralian Subsystem. The series spans the Asselian and Sakmarian stages. Fusulinid assemblages of both

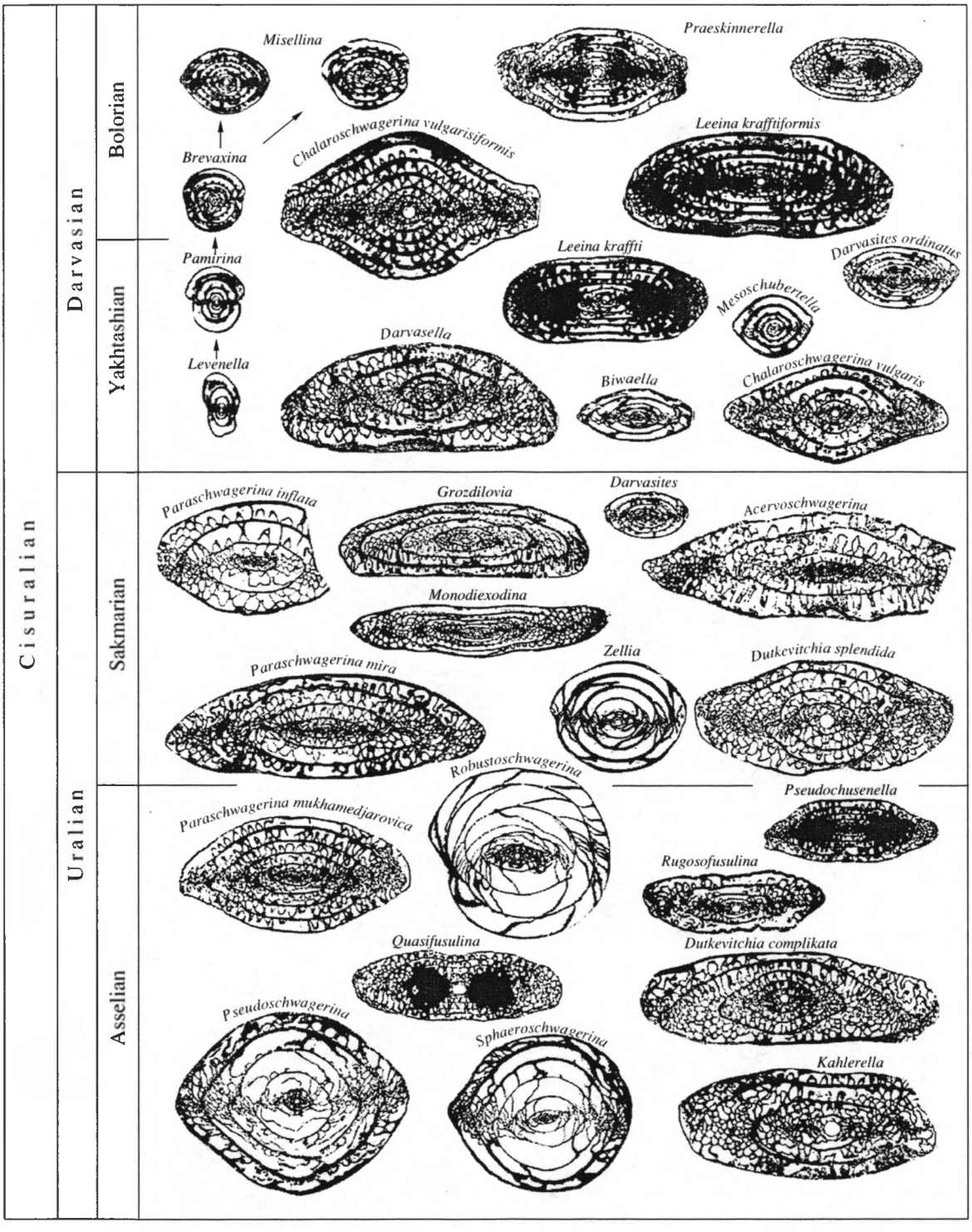

Plate I. Most characteristic fusulinid assemblages of series and stages of the Cisuralian Subsystem.

stages are similar and tightly connected with the Gzhelian ones. The order Schwagerinida is dominant. Genera *Rugosofusulina, Dutkevitchia, Kahlerella* (family Rugosofusulinidae), and *Pseudofusulina* (family Pseudofusulinidae), which appeared in the Gzhelian Age, are among the most abundant and characteristic. Especially typical are genera with swollen and loosely coiled shells, such as *Sphaeroschwagerina, Pseudoschwagerina, Paraschwagerina, Zellia, Robustoschwagerina*, and some others, which are artificially attributed to the heterogeneous family Schwagerinidae. The genus *Quasifusulina* (order Fusulinida) is also

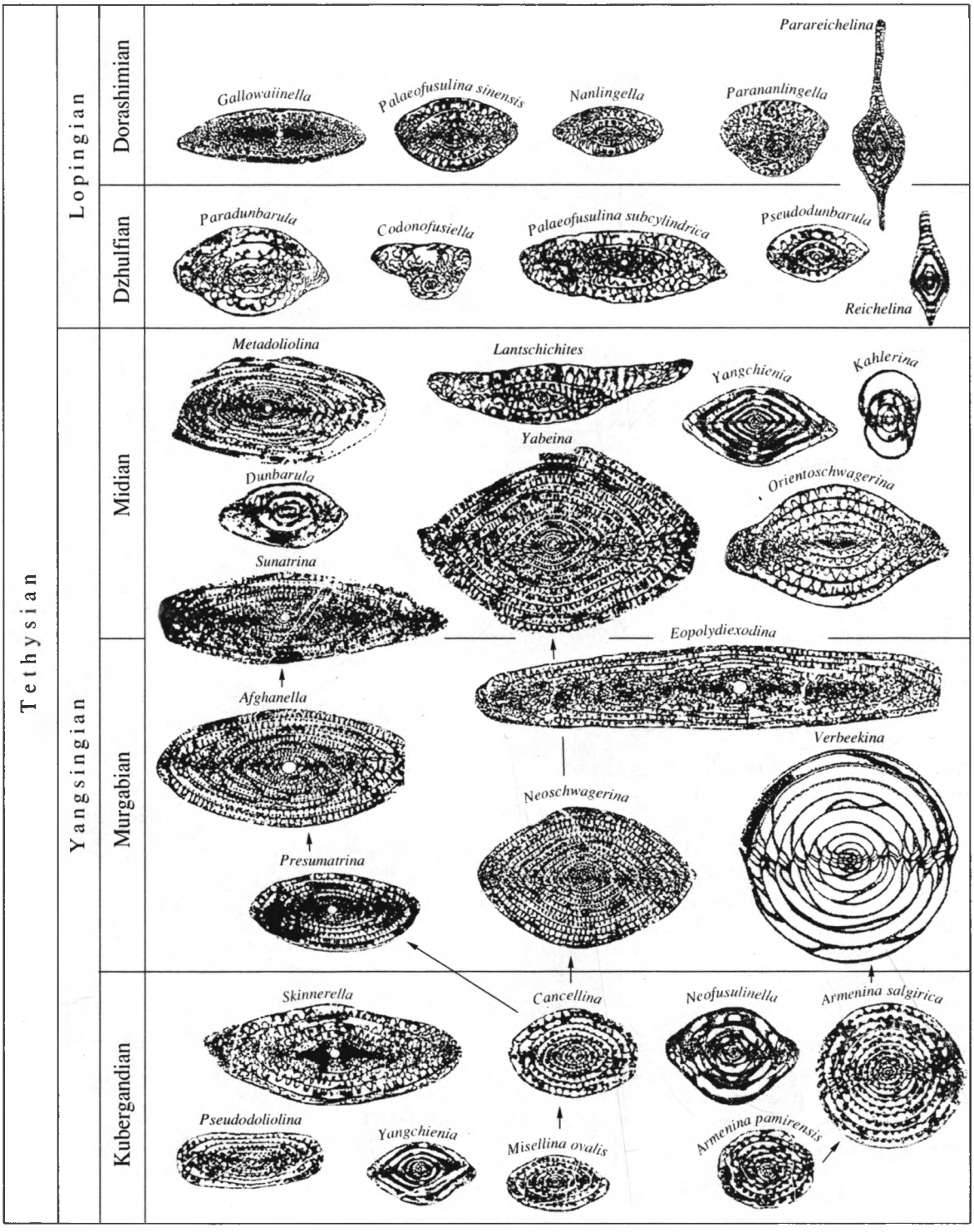

Plate II. Most characteristic fusulinid assemblages of series and stages of the Yangsingian Subsystem.

remarkable. Distinctions between Asselian and Sakmarian fusulinid assemblages are perceptible mostly at the species level, although the Sakmarian strata yield *Eoparafusulina* in association with first *Darvasites* and *Monodiexodina* forms, all unknown from the underlying deposits.

The Darvasian Series includes the Yakhtashian and Bolorian stages. Dutkevich (1937) introduced the series name to define an upper part of the Lower Permian in the Darvaz. Miklukho-Maclay (1958) distinguished under this name the stage corresponding to the *Parafusulina* and *Misellina* fusulinid zones. In addition

to *Misellina* Beds, the latter included strata with *Armenina* and first *Cancellina* forms of the Kubergandian Horizon. Later on, these strata were attributed to the individual Kubergandian Stage of the Upper Permian (Leven, 1963). Accordingly, the Darvasian Series accepted in this work is of a reduced range as compared to the primary one.

Fusulinid assemblages of the Yakhtashian and Bolorian stages of the series are very similar to each other, being clearly distinct from those of the previous Uralian Series. The boundary between two series marks extinction of most genera representing the family Schwagerinidae and a sharp decrease in abundance and diversity of the family Rugosofusulinidae. In the order Schwagerinida, most characteristic are *Chalaroschwagerina, Darvasites, Leeina* from the *L. kraffti–L. fusiformis* group, *Praeskinnerella,* and first *Parafusulina* and *Skinnerella.* Genera *Biwaella, Mesoschubertella,* and *Toriyamaia* are remarkable representatives of the other orders. *Pamirina* that appeared in the Yakhtashian Age is of a special importance. At the beginning of the Bolorian Age, this genus gave rise to the first higher fusulinids of the order Neoschwagerinida (genus *Misellina*). The Darvasian fusulinid assemblages differ in general from the Uralian ones despite their successive character. At the same time, they include first taxa of the later Yangsingian Series.

Tethysian Subsystem

The Yangsingian Series spans the Kubergandian, Murgabian, and Midian Stages. Huang (1932) who distinguished the series attributed to it a part of the carbonate succession in southern China: the Chihsia Formation beginning from its transgressive base and the entire Maokou Formation up to the base of the next transgressive complex of the Lopingian Series. In the recent stratigraphic schemes of China (Sheng and Jin, 1994), the series is divided into four stages and corresponds to the section interval from the *Misellina* Zone to the *Yabeina* Zone inclusive. It should be noted that *Misellina* forms from the base of the Chihsia type section are represented by highly developed species *M. claudiae* and *M. ovalis* characteristic of the Kubergandian Stage. Their occurrence is indicative of the lower series boundary, which I understand as corresponding to the base of the *Armenina–Misellina ovalis* Zone of the Kubergandian Stage. Across this boundary, there is recorded the most remarkable change in composition of Permian fusulinid communities. *Chalaroschwagerina, Darvasites, Leeina, Biwaella, Mesoschubertella, Toriyamaia,* last *Quasifusulina, Robustoschwagerina,* and *Acervoschwagerina* characteristic of the Darvasian Series completely disappear here. The fusulinid assemblage of the Kubergandian Stage is entirely renewed. Higher fusulinids of the order Neoschwagerinida, newcomers among which are genera *Armenina, Pseudodoliolina* and *Cancellina,* grow in abundance. The last two genera are ancestors of families Pseudodoliolinidae and Neoschwagerinidae, which became prosperous in the Murgabian and, especially, Midian ages together with representatives of Sumatrinidae and Verbeekinidae whose origin is recorded in the initial Murgabian time. A considerable renewal is characteristic of the order Schwagerinida, the leading taxa of which are genera *Parafusulina, Skinnerella,* and *Chusenella.* First *Eopolydiexodina* represented by subgenus *Bidiexodina* appear in the second half of the Kubergandian Age. The monotypic family Yanchienidae very characteristic of the entire Yangsingian Series appears in the order Schubertellida at the beginning of the Kubergandian Age. Generic composition of Schubertellidae and Boultoniidae becomes renewed because of *Neofusulinella, Codonofusiella,* and *Wutuella* origin in the Kubergandian Age and of *Russiella, Dunbarula,* and *Lantschichites* appearance in the Murgabian and Midian ages. Genera *Nankinella, Staffella,* and *Sphaerulina,* the newcomers of the order Staffellida, grow in abundance. Thus, the Yangsingian epoch in general is characterized by a quick diversification and prosperity of the order Neoschwagerinida at the beginning of the Kubergandian Age, by appearance of highly organized multiapertural Schwagerinida representatives (genus *Eopolydiexodina*), and by the growing diversity of other fusulinid orders.

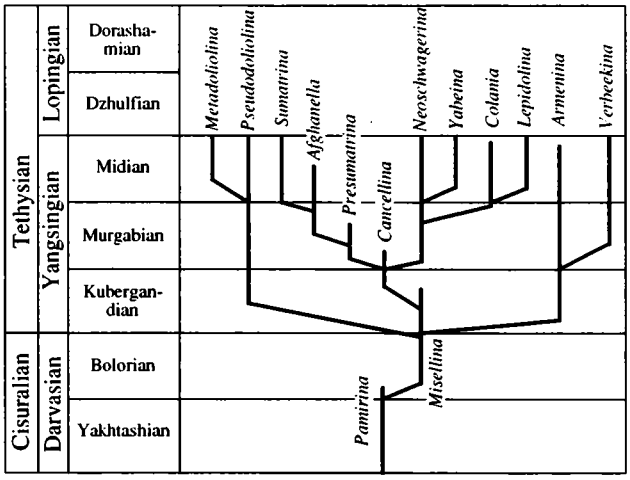

Fig. 4. Scheme illustrating phylogeny of the order Neoschwagerinida.

The Lopingian Series includes the Dzhulfian and Dorashamian stages. Richthofen used the name "Lopingian" to determine the Upper Permian coal-bearing deposits in China (Kayser,1883). Huang (1932) also considered these deposits in the rank of the upper series of the Permian System.

Fusulinid assemblages of the Lopingian Series are less diverse than the older ones because of a catastrophic extinction (71% of all genera) by the end of the Midian Age. Orders Neoschwagerinida and Schwagerinida completely disappeared across the Midian–Dzhulfian boundary, when families Yangchieniidae,

Schubertellidae, and Kahlerinidae became extinct in other orders. Three orders Staffellida, Ozawainellida, and Schubertellida crossed, however, this boundary. The former two became less diverse therewith, while families Palaeofusulinidae and Boultoniidae of the order Schubertellida experienced a certain renewal and diversification because of appearance of genera *Palaeofusulina, Paradunbarula, Tewoella*, and *Gallowaiinella* in the Dzhulfian Age. Genera *Nanlingella* and *Parananlingella* appeared somewhat later.

Thus, the Lopingian epoch in general was a survival time of fusulinids, which escaped the late Midian crisis. Families Palaeofusulinidae and Boultoniidae, which survived the crisis, were still in process of development, but even they, like other fusulinids, became extinct by the end of the Permian.

THE PROBLEM OF BOUNDARIES BETWEEN SUBSYSTEMS AND SERIES

If fusulinids are regarded as an orthostratigraphic group, the boundary between subsystems should be placed between the Bolorian and Kubergandian stages, since this level marks the most essential and quick changes in fusulinid assemblages. In the Kubergandian Stage stratotype and hypostratotype (the southeastern Pamirs), this level is at the base of the *Armenina–Misellina ovalis* Zone (Leven, 1981; Chediya *et al.*, 1986). In the Transcaucasia, the appearance level of abundant *Armenina* and highly developed *Misellina* is at the base of the Asni Formation (Leven, 1998). In southern China, deposits corresponding to the Bolorian Stage coupled with a greater part of the Kubergandian Stage are attributed now to the Loudinian Stage (Sheng and Jin, 1994), and the boundary under consideration is therefore inside the latter. I had a good opportunity to realize it personally when studied collections of thin sections, which characterize several very complete sections of southern China and are stored at the Nanking Institute of Paleontology and Stratigraphy and at the Guiyang Geological Survey, Guizhou Province. In the Luodian section for instance, first fusulinids undoubtedly belonging to the Kubergandian Stage (*Armenina, Parafusulina, Yangchienia*) appear in the Bed 25 (Excursion Guidebook, 1994; Leven, 2001). Immediately below, deposits are barren of fusulinids. In the Bed 21b, they belong to the upper Bolorian Substage (*Misellina termieri*). Conodonts *Mesogondolella saraciniensis* Gullo et Kozur, *M. phosphoriensis* (Younquist, Hawley et Miller) and *M. siciliensis* (Kozur), which coexist with fusulinids in the Bed 25, have been attributed by Kozur (1998) to the Roadian Stage of the Guadalupian Series. In sections of the Guizhou, Yunnan, and Guangxi provinces, which yield fusulinid assemblages, the boundary in question separates beds with diverse *Misellina, Chalaroschwagerina*, and *Darvasites* of the *Misellina termieri* Zone from beds with first *Armenina, Pseudodoliolina, Parafusulina*, and *Yangchienia* used to be included into the *Misellina claudiae* or *M. ovalis* Zone (Xiao *et al.*, 1986; Zhou *et al.*, 1987; Zhang *et al.*, 1988). The Bolorian–Kubergandian boundary is also well traceable in the Akiyoshi section of Japan, where it corresponds to the base of the *M. claudiae* Zone (*Guidebook...*, 1990; Ueno,1996). In Thailand, Bed B_2 yielding first *Armenina* and *Pseudodoliolina* definitely belongs to the lower Kubergandian Substage, whereas fusulinids from Bed B_1 are of the transitional type (Toriyama, 1975).

The above data show that outlined changes in fusulinid assemblages crossing the Lower–Upper Permian boundary are persistent and easily traceable throughout the Tethyan realm. Outside the Tethys, it is not a simple task to establish the same, because fusulinids become extinct across this boundary in the East European and Boreal regions, and this event may clearly indicate the boundary significance. Being based on ammonoids and conodonts, the Lower–Upper Permian boundary in the traditional East European scale separates the Kungurian and Ufimian stages (Leonova and Dmitriev, 1989; Kozur, 1995, 1998; Kozur *et al.*, 1994).

Because of fusulinid endemism, it is difficult to correlate the Tethyan and North American sections. In the project of the Global Chronostratigraphic Scale (Jin *et al.*, 1997), the boundary under consideration is placed at the base of the Roadian Stage, i.e., at the base of the Guadalupian Series. Such a decision is mostly based on distribution of ammonoids, as the ammonoid genera *Epiglyphioceras, Stacheoceras*, and *Paraceltites*, which are typical of the Roadian Stage in North America, are also known from the base of the Kubergandian Stage of the stratotype area in the southeastern Pamirs (Chediya *et al.*, 1986). In addition, the Bolorian ammonoids and conodonts are comparable with those of the Cathedralian Stage underlying the Roadian one (Leonova and Dmitriev, 1989). The suggested correlation seems doubtful however, because Chinese researchers reported that conodont species *Jinogondolella nankingensis*, the appearance of which marks by definition, the lower boundary of the Roadian Stage and of the Guadalupian Series as well, do not occur below Bed 46 of the Luodian section (Guizhou Province), i.e., below the upper third of the Murgabian Stage (Henderson *et al.*, 1999). One should remember also that Roadian conodonts coexist with Kubergandian fusulinids in the Bed 25 situated 150 m below in the Luodian section (Kozur, 1998).

The indicated contradiction in definition of the Cisuralian–Guadalupian boundary and in its correlation with the Cisuralian–Tethysian boundary are lacking explanations so far, as it has been formerly discussed in detail (Leven, 2001). Here I should mention only that the idea of Henderson and his Chinese colleagues who placed the boundary separating series or subsystems inside the Murgabian Stage is unacceptable in terms of fusulinids' evolution. Their idea excessively increases the range of the Cisuralian Subsystems and of

the upper stage of the Darvasian Series. Hence, it would be premature to use the name "Guadalupian" for the series spanning the Kubergandian, Murgabian, and Midian stages of the Tethyan scale.

Despite a considerable distinction between fusulinid assemblages of Uralian and Darvasian series of the Cisuralian Subsystem, it is difficult to establish precisely the level separating them. Since the changes are not abrupt but slightly progressing with time, the problem decision depends on formal criteria, which can be selected to fix the boundary. In Tethyan sections, the most convenient criterion is the appearance level of the *Pamirina* and approximately concurrent *Chalaroschwagerina* genera. With this criterion, the boundary coincides with the base of the Yakhtashian Stage (Leven, 19806) and can be traced easily from the Carnic Alps on the west (Forke, 1995) to Japan on the east (*Guidebook...*, 1990; Ueno, 1996). The most clear transition from Sakmarian to Yakhtashian fusulinid assemblages is observable in the South Chinese sections, where it is recorded across boundary between the Zisongian and Longlinian regional stages (Sheng and Jin, 1994), the former of which spans the Asselian and Sakmarian stages, whereas the latter one is a younger synonym of the Yakhtashian Stage.

In opinion of Leonova (see in Leven *et al.*, 1992), ammonoids found near the top of the Yakhtashian Stage in its stratotype area are similar to the late Artinskian counterparts. The lower Artinskian conodonts (*Sweetognathus whitei*) appear in the section almost concurrently with first *Pamirina* forms. Accordingly, we may assume that Yakhtashian and Artinskian stages are approximately correlative, and hence the boundary between Uralian and Darvasian series of the Cisuralian Subsystem should be at the level separating the Sakmarian and Artinskian stages of the East European scale. Nothing more definite can be said about the issue based on the data available at present. Moreover, the boundary between these stages has not been officially fixed and present opinions about its positioning are very diverse. In North America, it used to be regarded as correlative with the boundary between the Lenoxian and Hessian stages (Ross and Ross, 1987), but its position is probably lower, inside the Lenoxian Stage or even at the base of the latter (Wardlaw and Davydov, 2000). The problem can be solved after additional research.

There is no problem concerning the boundary between the Yangsingian and Lopingian series, because it marks a catastrophic extinction of fusulinids and many other marine invertebrates by the end of the Midian Age (Jin, 1993). The GSSP of this boundary is suggested to be either at the base of the *Clarkina postbitteri* Zone in the conodont scale, or slightly higher, at the base of the *C. dukouensis* Zone (Henderson *et al.*, 2000). In the East European scale, its position is still unclear. Taking into consideration the paleomagnetic dates, one should seek for it inside the upper Tatarian Substage (Burov, 1996), because the boundary between the Kiama and Illawarra superchrons corresponds here to the substage base, whereas in the Tethys it is near the base of the Midian Stage (Jin *et al.*, 2000).

STAGES AND THEIR BOUNDARIES

Except for two upper stages, all the others are defined in the Tethyan scale based on fusulinids. General characteristics of fusulinid assemblages have been used here to distinguish the Asselian and Sakmarian stages, while the other ones correspond to stages in evolution of fusulinids of the *Pamirina–Misellina–Cancellina–Neoschwagerina–Yabeina* phylogenetic lineage (Fig. 4, Plates I, II).

The Asselian Stage can be characterized based on fusulinid assemblages known from its type sections in the southern Urals. Most remarkable and typical of the unit are loosely coiled fusulinids, such as *Sphaeroschwagerina, Pseudoschwagerina, Zellia, Robustoschwagerina*, and some others. *Quasifusulina, Rugosofusulina, Dutkevitchia, Kahlerella*, and *Pseudofusulina* are also characteristic and abundant. Compositional peculiarities of Asselian fusulinid assemblages are persistent, though with some variations, throughout the Tethyan realm, and the stage is recognizable without difficulties everywhere from the Alps on the west to Japan and Indochina on the east.

The Sakmarian Stage yields fusulinids of a similar composition and can be discriminated from the Asselian Stage not everywhere. For instance, both subdivisions are regarded as the integral Zisongian Stage in stratigraphic schemes recently accepted in China (Sheng and Jin, 1994). Nevertheless, certain distinctions, even at the generic level, become evident after a careful analysis. Hence, the relative abundance of aforementioned Asselian genera, all occurring in the Sakmarian Stage as well, is different: *Quasifusulina, Sphaeroschwagerina, Pseudoschwagerina*, and *Dutkevitchia* are less abundant in the last case, while *Robustoschwagerina, Paraschwagerina*, and *Zellia* turn into leading taxa. Genera *Darvasites, Eoparafusulina, Monodiexodina* and *Acervoschwagerina* appear for the first time and become widespread. *Rugosofusulina* and *Pseudofusulina* are as abundant as before, but their species composition is considerably renewed. All these characteristics enable discrimination of the Sakmarian Stage, although its boundary with the Asselian Stage remains vague, especially if we take into consideration that it is imprecisely defined in the type sections of the southern Urals (Wardlaw *et al.*, 1999).

The Yakhtashian Stage is initial one in the Darvasian Series of the Cisuralian Subsystem. As is mentioned above, fusulinids of the stage differ from their Sakmarian assemblages as a consequence of the post-Sakmarian extinction, which decreased abundance rates of many genera typical of the Asselian–Sakmarian

community, namely of *Dutkevitchia, Sphaeroschwagerina, Pseudoschwagerina, Zellia, Paraschwagerina,* and some others. Newcomers characteristic of the Yakhtashian and Bolorian stages are genera *Pamirina, Pseudoreichelina, Mesoschubertella, Toriyamaia, Chalaroschwagerina, Darvasella, Leeina,* and *Praeskirnnerella. Darvasites* forms are still abundant like in the Sakmarian Stage, but their shells have more massive and wide chomata resembling those of the genus *Nagatoella.*

Deposits bearing Yakhtashian fusulinids are established in Turkey, Afghanistan, Darvaz, Pamirs, and Karakorum (Leven, 1967, 1980b, 1995, 1997; Leven *et al.*, 1992; Gaetani *et al.*, 1995). In China, the same interval of sedimentary successions corresponds to the Longlinian Stage (Sheng and Jin, 1994). The Yakhtashian Stage is also discriminated in the Akiyoshi section of Japan (*Guidebook...*, 1990; Ueno, 1996). Fusulinids characteristic of the stage are known from Thailand as well (Sakagami and Iwai, 1974).

The lower limitation of the Yakhtashian Stage has been discussed already in the section devoted to boundaries, which separate series of the Cisuralian Subsystem. It is problematic so far. At present, the boundary in question can be defined based on concurrent appearance of *Chalaroschwagerina* and first *Pamirina* forms of the subgenus *P. (Levenella).* The Yakhtashian Stage in my understanding approximately corresponds to the Artinskian Stage of the East European scale, as one can judge from distribution of ammonoids and conodonts. Because of biogeographic factors, a comparison between fusulinid assemblages from the Tethys and the Urals neither proves, nor discards that inference. Some Yakhtashian genera known from Texas (*Chalaroschwagerina, Praeskinnerella*) suggest that the unit under consideration may be correlative with the Lenoxian Stage (or with its upper substage) coupled with a basal interval of the Hessian Stage. Data on conodonts (Wardlaw and Davydov, 2000) are consistent with this conclusion. On the other hand, the Lenoxian Stage seem to be correlative with the Sakmarian one based on distribution of ammonoids (Kotlyar *et al.*, 1987).

The Bolorian Stage yields fusulinid assemblages, which are similar in general to the Yakhtashian counterparts. Distinctions are mainly recognizable at the species level. The principal distinctive feature is the appearance and subsequent flourishing of primitive higher fusulinids of the order Neoschwagerinida, which were represented first by subgenus *Brevaxina,* and then by subgenus *Misellina* (*Misellina*) (Fig. 4). A quick branching of *Brevaxina* from the ancestral genus *Pamirina* (Fig. 6) is recorded in many sections, and, based on this event, the Yakhtashian–Bolorian boundary is easily and confidently recognizable in all areas of the Tethys. Coexistence of *Brevaxina* and primitive *Misellina* with Yakhtashian–Bolorian genera *Mesoschubertella, Toriyamaia, Chalaroschwagerina, Darvasites, Leeina,* and *Praeskinnerella* enabled recognition of the Bolorian Stage in many sections of the Transcaucasia and Iran on the west to Japan and Indochina on the east. The lower part of the Luodinian Stage in China (*Brevaxina dyhrenfurthi* and *Misellina termieri* zones) is correlative with the stage as well (Sheng and Jin, 1994).

Outside the Tethys, the Bolorian Stage is unrecognizable because of endemism of its fusulinid assemblages. Based on position in the sections and on distribution of ammonoids and conodonts, the unit is correlative with the Kungurian Stage of Eastern Europe and with the Cathedralian Stage of North America (Leonova and Dmitriev, 1989; Kozur, 1995, 1998; Kozur *et al.*, 1994). The correlation is problematic so far.

The Kubergandian Stage differs strongly from the Bolorian Stage and yields contrasting fusulinid assemblages. The distinctions have been outlined above by consideration of fusulinid evolution and boundaries separating the Permian Subsystems. Here, I should mention only that fusulinids from the lower *Armenina–Misellina ovalis* Zone of the stage are tightly connected with Bolorian assemblages, because the genus *Misellina* is well represented at this level. Accordingly, some experts, for instance, my colleagues in China, feel it appropriate to couple the Bolorian Stage and that zone into one stratigraphic subdivision such as the Luodinian regional stage (Sheng and Jin, 1994). However, the genus *Misellina* is represented in the *Armenina–M. ovalis* Zone by species, which are more developed, than their Bolorian counterparts. In addition, exactly this level marks the initial renewal of the entire community, i.e., the appearance of taxa, such as *Armenina, Pseudodoliolina, Yangchienia,* and *Neofusulinella,* which become widespread in the Yangsingian Series (Figs. 5, 6). The majority of taxa characteristic of the Yakhtashian–Bolorian period in fusulinid evolution do not cross the zone lower boundary. This is established not only in type sections of the Kubergandian Stage in the Pamirs (Leven, 1981; Chediya *et al.*, 1986), but also in most complete sections of southern China, where renewals in assemblages are recorded either at the base, or inside the beds corresponding to the *Misellina claudiae* or *M. ovalis* zones (Xiao *et al.*, 1986; Zhou *et al.*, 1987).

Deposits with Kubergandian fusulinids are established in numerous sections of a vast region, from the Mediterranean on the west to Japan and Indochina on the east. At present, there is no data, which may substantiate the direct correlation between the Kubergandian Stage and subdivisions of the East European scale. I accept conventionally the correlation between lower boundaries of this and Ufimian stages. My comments to correlation of the Kubergandian and Guadalupian lower boundaries are presented above.

The Murgabian Stage is tightly connected with the Kubergandian Stage as far as it concerns their fusulinid assemblages. Diversification of the order Neoschwagerinida is in progress at this level. Family Sumatrinidae with genera *Presumatrina, Afghanella,* and *Sumat-*

rina, Verbeekinidae (genus *Verbeekina*), and subfamily Lepidolininae (genera *Gifuella* and *Colania*) represent taxa first appeared in that order (Fig. 4). Genera *Yangchienia, Chusenella, Skinnerella*, and *Eopolydiexodina* of the Kubergandian origin become widespread. In distinction from the Kubergandian Age, the last taxon is represented by subgenus *E. (Eopolydiexodina)*. Orders Schubertellida and Staffellida also experience diversification.

The stage lower boundary is placed by definition at the appearance level of *Neoschwagerina, Presumatrina*, and *Verbeekina*, i.e., at the base of the *N. simplex* Zone. This is a very significant and remarkable level, and one should place the stage boundary exactly here, but not at the lower appearance level of the genus *Cancellina*, as it is done in some works by Chinese researchers (Sheng and Jin, 1994; Zhu and Zhang, 1994).

The Murgabian Stage and its lower boundary are easily traceable based on fusulinids from the Mediterranean to Indochina, i.e., over the entire Tethyan region. It is impossible so far to correlate directly the stage with subdivisions of the East European scale. As for the stage correlation with the North American scale, it seem to be very problematic because of controversial interpretations of conodont dates (Kozur, 1998; Henderson *et al.*, 1999; Leven, 2001).

The Midian Stage marks the third peak of fusulinid generic diversity (Figs. 1, 2). It was a prosperity time of the order Neoschwagerinida represented by taxa, which were highly organized, crowning the phylogenetic lineages of the Neoschwagerinidae, Sumatrinidae, Verbeekinidae, and Pseudodoliolinidae families (Fig. 4). Most remarkable among these taxa were genera *Yabeina* and *Lepidolina* first appeared at this level. Subfamily Chusenellinae with characteristic new genus *Orientoschwagerina* was of a greater significance in the order Schwagerinida. The Midian Age signifies as well the peak development of the orders Schubertellida, Staffellida, and Ozawainellida. Most remarkable representatives of these orders are the first appeared genera *Lantschichites, Paradoxiella, Eoverbeekina, Reichelina*, and *Rauserella*. One more taxon that appeared and became widespread in the Midian Age is family Kahlerinidae whose genetic roots are unclear.

According to original definition, the lower boundary of Midian Stage corresponds to the first occurrence level of *Yabeina* and *Lepidolina* whose geographic range is restricted by eastern areas of the Tethys, beyond which the boundary can be traced not so easily. Because of various criteria used to define the boundary on the west, it looks somewhat blurred here (Leven, 1996). Certainly, the boundary proper and its distinctive characters should be clarified and coordinated in the future. Paleomagnetic data may be important in this case, because preliminary (not very confident) research showed that boundary between the Kiama and Illawarra superchrons is confined to the Midian Stage base (Jin *et al.*, 2000).

With certain reservations concerning the lower boundary, the Midian Stage is readily recognizable in the Tethyan realm. In China, it approximately corresponds to the Lengwuan Stage (Sheng and Jin, 1994). In East European sections, the Kiama–Illawarra boundary is detected inside the Tatarian Stage, and the Midian Stage may correspond therefore to the basal interval of the upper Tatarian Substage. In North America, the stage under consideration (or its greater upper part) is likely correlative with the Capitanian Stage. This is evident from the fact that conodonts *Jinogondolella postserrata*, the appearance of which marks the lower boundary of the Capitanian Stage, coexist with Midian fusulinids in the upper part of Maokou Limestone in China (Mei *et al.*, 1998). Typical Midian fusulinids (*Yabeina, Reichelina, Paradoxiella*) are known from the Lamar Limestone crowning the Capitanian Formation of Texas (Skinner and Wilde, 1955).

The Dzhulfian Stage is clearly distinct from the Midian Stage, because its fusulinid assemblages are of a low diversity. They are lacking fusulinids of the orders Schwagerinida and Neoschwagerinida, which dominated before and became extinct by the end of the Midian Age (Plate II). The other orders lost families Yangchieniidae, Schubertellidae, and Kahlerinidae. Families Boultoniidae and Palaeofusulinidae survived the late Midian crisis and, experiencing diversification, gave rise to new genera *Gallowaiinella, Tewoella, Paradunbarula*, and *Palaeofusulina*, which are most highly organized in corresponding families. Three of these genera used to be regarded not long ago as characteristic of the Dorashamian Stage only, but recently they were found in beds of southern China, which correspond to the lower part of the Dzhulfian Stage (Zhu, 1996).

The Dzhulfian Stage lower boundary (or boundary between the Yangsingian and Lopingian series), which is well detectable by changes in fusulinid assemblages, is now under verification based on distribution of conodonts (Henderson *et al.*, 2000). The GSSP of that boundary is supposed to be at the base of the Wuchiapingian Stage in the Penglaiten section of southern China (Henderson *et al.*, 2000). According to recent data (Sweet and Mei, 1999), the same level is recognizable at the base of the Ali Bashi Formation in the Kuhi Ali-Bashi section of northwestern Iran. The formation approximately corresponds to the lower part of *Reichelina–Codonofusiella* Beds in sections of Transcaucasia (Leven, 1998) and central Iran (*Iranian-Japanese...*, 1981). In North America, the boundary coincides with the top of the Capitanian Stage.

The Dorashamian Stage is unclearly characterized in terms of fusulinids' distribution. Characteristic of the stage are genera *Palaeofusulina, Nanlingella, Parananlinhella, Paradunbarula, Gallowaiinella, Tewoella, Nankinella, Sphaerulina, Staffella, Reiche-*

community, namely of *Dutkevitchia, Sphaeroschwagerina, Pseudoschwagerina, Zellia, Paraschwagerina,* and some others. Newcomers characteristic of the Yakhtashian and Bolorian stages are genera *Pamirina, Pseudoreichelina, Mesoschubertella, Toriyamaia, Chalaroschwagerina, Darvasella, Leeina,* and *Praeskirnnerella. Darvasites* forms are still abundant like in the Sakmarian Stage, but their shells have more massive and wide chomata resembling those of the genus *Nagatoella.*

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The lower limitation of the Yakhtashian Stage has been discussed already in the section devoted to boundaries, which separate series of the Cisuralian Subsystem. It is problematic so far. At present, the boundary in question can be defined based on concurrent appearance of *Chalaroschwagerina* and first *Pamirina* forms of the subgenus *P. (Levenella).* The Yakhtashian Stage in my understanding approximately corresponds to the Artinskian Stage of the East European scale, as one can judge from distribution of ammonoids and conodonts. Because of biogeographic factors, a comparison between fusulinid assemblages from the Tethys and the Urals neither proves, nor discards that inference. Some Yakhtashian genera known from Texas (*Chalaroschwagerina, Praeskinnerella*) suggest that the unit under consideration may be correlative with the Lenoxian Stage (or with its upper substage) coupled with a basal interval of the Hessian Stage. Data on conodonts (Wardlaw and Davydov, 2000) are consistent with this conclusion. On the other hand, the Lenoxian Stage seem to be correlative with the Sakmarian one based on distribution of ammonoids (Kotlyar *et al.*, 1987).

The Bolorian Stage yields fusulinid assemblages, which are similar in general to the Yakhtashian counterparts. Distinctions are mainly recognizable at the species level. The principal distinctive feature is the appearance and subsequent flourishing of primitive higher fusulinids of the order Neoschwagerinida, which were represented first by subgenus *Brevaxina*, and then by subgenus *Misellina* (*Misellina*) (Fig. 4). A quick branching of *Brevaxina* from the ancestral genus *Pamirina* (Fig. 6) is recorded in many sections, and, based on this event, the Yakhtashian–Bolorian boundary is easily and confidently recognizable in all areas of the Tethys. Coexistence of *Brevaxina* and primitive *Misellina* with Yakhtashian–Bolorian genera *Mesoschubertella, Toriyamaia, Chalaroschwagerina, Darvasites, Leeina,* and *Praeskinnerella* enabled recognition of the Bolorian Stage in many sections of the Transcaucasia and Iran on the west to Japan and Indochina on the east. The lower part of the Luodinian Stage in China (*Brevaxina dyhrenfurthi* and *Misellina termieri* zones) is correlative with the stage as well (Sheng and Jin, 1994).

Outside the Tethys, the Bolorian Stage is unrecognizable because of endemism of its fusulinid assemblages. Based on position in the sections and on distribution of ammonoids and conodonts, the unit is correlative with the Kungurian Stage of Eastern Europe and with the Cathedralian Stage of North America (Leonova and Dmitriev, 1989; Kozur, 1995, 1998; Kozur *et al.*, 1994). The correlation is problematic so far.

The Kubergandian Stage differs strongly from the Bolorian Stage and yields contrasting fusulinid assemblages. The distinctions have been outlined above by consideration of fusulinid evolution and boundaries separating the Permian Subsystems. Here, I should mention only that fusulinids from the lower *Armenina–Misellina ovalis* Zone of the stage are tightly connected with Bolorian assemblages, because the genus *Misellina* is well represented at this level. Accordingly, some experts, for instance, my colleagues in China, feel it appropriate to couple the Bolorian Stage and that zone into one stratigraphic subdivision such as the Luodinian regional stage (Sheng and Jin, 1994). However, the genus *Misellina* is represented in the *Armenina–M. ovalis* Zone by species, which are more developed, than their Bolorian counterparts. In addition, exactly this level marks the initial renewal of the entire community, i.e., the appearance of taxa, such as *Armenina, Pseudodoliolina, Yangchienia,* and *Neofusulinella,* which become widespread in the Yangsingian Series (Figs. 5, 6). The majority of taxa characteristic of the Yakhtashian–Bolorian period in fusulinid evolution do not cross the zone lower boundary. This is established not only in type sections of the Kubergandian Stage in the Pamirs (Leven, 1981; Chediya *et al.*, 1986), but also in most complete sections of southern China, where renewals in assemblages are recorded either at the base, or inside the beds corresponding to the *Misellina claudiae* or *M. ovalis* zones (Xiao *et al.*, 1986; Zhou *et al.*, 1987).

Deposits with Kubergandian fusulinids are established in numerous sections of a vast region, from the Mediterranean on the west to Japan and Indochina on the east. At present, there is no data, which may substantiate the direct correlation between the Kubergandian Stage and subdivisions of the East European scale. I accept conventionally the correlation between lower boundaries of this and Ufimian stages. My comments to correlation of the Kubergandian and Guadalupian lower boundaries are presented above.

The Murgabian Stage is tightly connected with the Kubergandian Stage as far as it concerns their fusulinid assemblages. Diversification of the order Neoschwagerinida is in progress at this level. Family Sumatrinidae with genera *Presumatrina, Afghanella,* and *Sumat-*

rina, Verbeekinidae (genus *Verbeekina*), and subfamily Lepidolininae (genera *Gifuella* and *Colania*) represent taxa first appeared in that order (Fig. 4). Genera *Yangchienia, Chusenella, Skinnerella*, and *Eopolydiexodina* of the Kubergandian origin become widespread. In distinction from the Kubergandian Age, the last taxon is represented by subgenus *E. (Eopolydiexodina)*. Orders Schubertellida and Staffellida also experience diversification.

The stage lower boundary is placed by definition at the appearance level of *Neoschwagerina, Presumatrina*, and *Verbeekina*, i.e., at the base of the *N. simplex* Zone. This is a very significant and remarkable level, and one should place the stage boundary exactly here, but not at the lower appearance level of the genus *Cancellina*, as it is done in some works by Chinese researchers (Sheng and Jin, 1994; Zhu and Zhang, 1994).

The Murgabian Stage and its lower boundary are easily traceable based on fusulinids from the Mediterranean to Indochina, i.e., over the entire Tethyan region. It is impossible so far to correlate directly the stage with subdivisions of the East European scale. As for the stage correlation with the North American scale, it seem to be very problematic because of controversial interpretations of conodont dates (Kozur, 1998; Henderson *et al.*, 1999; Leven, 2001).

The Midian Stage marks the third peak of fusulinid generic diversity (Figs. 1, 2). It was a prosperity time of the order Neoschwagerinida represented by taxa, which were highly organized, crowning the phylogenetic lineages of the Neoschwagerinidae, Sumatrinidae, Verbeekinidae, and Pseudodoliolinidae families (Fig. 4). Most remarkable among these taxa were genera *Yabeina* and *Lepidolina* first appeared at this level. Subfamily Chusenellinae with characteristic new genus *Orientoschwagerina* was of a greater significance in the order Schwagerinida. The Midian Age signifies as well the peak development of the orders Schubertellida, Staffellida, and Ozawainellida. Most remarkable representatives of these orders are the first appeared genera *Lantschichites, Paradoxiella, Eoverbeekina, Reichelina*, and *Rauserella*. One more taxon that appeared and became widespread in the Midian Age is family Kahlerinidae whose genetic roots are unclear.

According to original definition, the lower boundary of Midian Stage corresponds to the first occurrence level of *Yabeina* and *Lepidolina* whose geographic range is restricted by eastern areas of the Tethys, beyond which the boundary can be traced not so easily. Because of various criteria used to define the boundary on the west, it looks somewhat blurred here (Leven, 1996). Certainly, the boundary proper and its distinctive characters should be clarified and coordinated in the future. Paleomagnetic data may be important in this case, because preliminary (not very confident) research showed that boundary between the Kiama and Illawarra superchrons is confined to the Midian Stage base (Jin *et al.*, 2000).

With certain reservations concerning the lower boundary, the Midian Stage is readily recognizable in the Tethyan realm. In China, it approximately corresponds to the Lengwuan Stage (Sheng and Jin, 1994). In East European sections, the Kiama–Illawarra boundary is detected inside the Tatarian Stage, and the Midian Stage may correspond therefore to the basal interval of the upper Tatarian Substage. In North America, the stage under consideration (or its greater upper part) is likely correlative with the Capitanian Stage. This is evident from the fact that conodonts *Jinogondolella postserrata*, the appearance of which marks the lower boundary of the Capitanian Stage, coexist with Midian fusulinids in the upper part of Maokou Limestone in China (Mei *et al.*, 1998). Typical Midian fusulinids (*Yabeina, Reichelina, Paradoxiella*) are known from the Lamar Limestone crowning the Capitanian Formation of Texas (Skinner and Wilde, 1955).

The Dzhulfian Stage is clearly distinct from the Midian Stage, because its fusulinid assemblages are of a low diversity. They are lacking fusulinids of the orders Schwagerinida and Neoschwagerinida, which dominated before and became extinct by the end of the Midian Age (Plate II). The other orders lost families Yangchieniidae, Schubertellidae, and Kahlerinidae. Families Boultoniidae and Palaeofusulinidae survived the late Midian crisis and, experiencing diversification, gave rise to new genera *Gallowaiinella, Tewoella, Paradunbarula*, and *Palaeofusulina*, which are most highly organized in corresponding families. Three of these genera used to be regarded not long ago as characteristic of the Dorashamian Stage only, but recently they were found in beds of southern China, which correspond to the lower part of the Dzhulfian Stage (Zhu, 1996).

The Dzhulfian Stage lower boundary (or boundary between the Yangsingian and Lopingian series), which is well detectable by changes in fusulinid assemblages, is now under verification based on distribution of conodonts (Henderson *et al.*, 2000). The GSSP of that boundary is supposed to be at the base of the Wuchiapingian Stage in the Penglaiten section of southern China (Henderson *et al.*, 2000). According to recent data (Sweet and Mei, 1999), the same level is recognizable at the base of the Ali Bashi Formation in the Kuhi Ali-Bashi section of northwestern Iran. The formation approximately corresponds to the lower part of *Reichelina–Codonofusiella* Beds in sections of Transcaucasia (Leven, 1998) and central Iran (*Iranian-Japanese...*, 1981). In North America, the boundary coincides with the top of the Capitanian Stage.

The Dorashamian Stage is unclearly characterized in terms of fusulinids' distribution. Characteristic of the stage are genera *Palaeofusulina, Nanlingella, Parananlinhella, Paradunbarula, Gallowaiinella, Tewoella, Nankinella, Sphaerulina, Staffella, Reiche-*

lina, and *Parareichelina*. Except for *Parananlingella* and *Parareichelina*, all other genera listed above are known also from the underlying Dzhulfian Stage. Accordingly, it is difficult to discriminate the Dzhulfian and Dorashamian stages based on fusulinids, but the problem can probably be solved with the help of conodonts and ammonoids (Jin *et al.*, 1997).

Outside southern China, where the local Changhsingian Stage is an equivalent of the Dorashamian Stage, fusulinids of the latter are well represented only in the Nikitinskaya and Urushten formations of the northern Caucasus (Kotlyar *et al.*, 1983). Based on conodonts, the stage is distinguishable in Transcaucasia (Kozur *et al.*, 1978) and in northwestern and central Iran (Sweet and Mei, 1999; Kozur *et al.*, 1975). The known finds of fusulinids and conodonts imply also that Dorashamian deposits are present in Greece, Turkey, Pamirs, the Salt Range, and Japan. Outside the Tethyan realm, deposits of the stage have not been established so far.

CONCLUSION

(1) The global chronostratigraphic scale is hardly applicable for Permian deposits in Tethyan areas, for which the alternative scale is suggested and characterized with due account for most recent data. The Permian System is divided therewith in two subsystems with two series in each. In general, these subdivisions and nine discriminated stages correspond to natural periods in evolution of fusulinids and other orthostratigraphic groups of marine fossils (Leven *et al.*, 1996). They reflect historical periods of the basin development and its biota evolution that progressed under influence of large transgressions and regressions, which increased or decreased the taxonomic diversity of organisms (Leven, 1993).

(2) The necessity to divide the Permian in two subsystems and four series is also implied by data on macrofauna distribution in Boreal sections. Accordingly, biostratigraphic regularities used to substantiate the Tethyan scale are of a global significance, being valid outside the Tethyan realm.

(3) The three-series global chronostratigraphic scale authorized recently is inconsistent with many biostratigraphic evidences considered in this work. Elaborators of the scale ignored the principle of natural history using instead the formal definition of stratigraphic boundaries based on conodonts' succession in a single phylogenetic lineage. As a result, stratigraphic ranges of series are disproportional in that scale and some boundaries, especially those of the Guadalupian Series and its stages, are traceable worldwide with difficulties. The series ranges in the scale proposed are more proportional, and the scale itself inherits advantages of the traditional two-part scale. The described subdivision scheme seems to be a better candidate for the global scale.

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Valanginian–Lower Hauterivian Brachiopod Assemblages from the Kacha–Bodrak Interfluve, Southwestern Crimea

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Abstract—Brachiopods are among the fossils groups frequently occurring in sections of the southwestern Crimea. The analysis of joint stratigraphic distribution of ammonite and brachiopod assemblages that has been stimulated by revision of the Lower Cretaceous stratigraphy in that area (Baraboshkin, 1997a) shows that brachiopods are suitable for the correct age determination of their host rocks. The following brachiopod zones and beds are discriminated in the Valanginian–lower Hauterivian of the Crimean Mountains: *Loriolithyris carteroniana* (lower Valanginian), *Loriolithyris metaitensis* (lower–upper Valanginian), *Musculina sanctaecrucis* (lower Hauterivian) with the lower *Musculina globus* and upper *Symphythyris neocomiensis* subzones, and the *Cyclothyris corallina neocomiensis* Zone, the latter distinguishable in shallow-water carbonate (reefal) and terrigenous facies. Some of these units are also established in other regions of the world. The new brachiopod species *Musculina nuciformis* Smirnova sp. nov., *M. nikischini* Smirnova sp. nov., *Terebrataliopsis semicircularis* Smirnova sp. nov., and *T. compacta* Smirnova sp. nov. are described.

Key words: Lower Cretaceous, Valanginian, Hauterivian, stratigraphy, brachiopods, ammonites, southwestern Crimea.

INTRODUCTION

The information on the Lower Cretaceous brachiopods from the southwestern Crimea has been reported by Eichwald (1865–1868), Loriol (1896), Karakash (1907), Smirnova (1972, 1990), and others. Recently, Smirnova (2001) published a review of these papers.

The Valanginian and lower Hauterivian brachiopods of the southwestern Crimea are more abundant than elsewhere in Eastern Europe. They are represented by 23 species representing ten genera, five families, and two orders. Many of the species are also known in Western Europe, where they are considered as the Valanginian–Hauterivian taxa. Brachiopods sampled anew in the bed-by-bed mode from the Valanginian and lower Hauterivian sediments and their thorough correlation with the ammonite zonation refine considerably stratigraphic distribution of these fossils. The ammonite biostratigraphy elaborated by Baraboshkin (Baraboshkin 1997a, b, 2001; Baraboshkin and Yanin, 1997; Baraboshkin, 1999) is partly revised in this paper (Table 1). Ammonites were studied previously by Baraboshkin and Mikhailova (1994, 2000).

The studied brachiopods have been collected in the Kacha River basin near the Verkhorech'e Village, and in the Kacha–Bodrak interfluve. We also used description of sections and fossils collected here by Baraboshkin in 1988–1997. In addition, we revised fossils collected by the Lower Cretaceous Field Group under guidance of V.V. Drushchits (the Chair of Paleontology, Moscow State University), which worked in the Crimea in 1954–1965, and by students of the Moscow State University who worked in the Prokhladnoe Village area during the periods of their geological practice. The entire collection No. 100 consisting of about 350 specimens is stored at the Museum of Earth Science, Moscow State University.

DESCRIPTION OF THE SECTIONS

The Valanginian–Hauterivian sediments that yielded most of the collected brachiopods, are well exposed at the foot of the Rezannaya and Belaya mountains on the right bank of the Kacha River. The following members are distinguishable here (from the base upward) (Fig. 1).

(I) Puddingstone–gravelstone member composed of three alternating rock varieties: brown calcareous, oolitic, sandy conglomerate (up to 0.2 m thick) bearing well-rounded quartz pebble with limonite crust; gray loose, clayey sandstone (up to 0.45 m thick) with scarce quartz pebble, limonite ooliths, and plant remains; and grayish green clay (up to 0.7 m thick). Large pebble and boulders of the underlying Tavrian Group occur at the base of the member, and a lens-like compact limestone bed up to 0.3 m thick is traceable in places. Each compact bed is separated from the underlying one by unconformity of the erosional or hardground (HG) type. The top of the member is ferruginate and burrowed, representing a thin (0.1–0.2 m) condensed horizon composed of *in situ* fragments of the eroded HG,

Table 1. Valanginian–Hauterivian biostratigraphic zonation, the Kacha–Bodrak interfluve

Stage	Substage	Standard zonation of the West Mediterranean Province (part); Hoedemaeker and Rawson, 2000		Kacha–Bodrak interfluve, Baraboshkin, 1997a	Kacha–Bodrak interfluve, this paper			
		Ammonite zone, subzone		Ammonite zones	Ammonite zones	Brachiopod zone, subzone		
Hauterivian	upper	*Pseudothurmannia angulicostata auctorum*	*Pseudothurmannia catulloi*	*Pseudothurmannia angulicostata*	*Pseudothurmannia angulicostata*	not studied		
			Pseudothurmannia angulicostata auct.					
		Balearites balearis		*Craspedodiscus discofalcatus*	*Milanowskia speetonensis*			
		Pseudospitidiscus ligatus		*Milanowskia speetonensis*	*Speetoniceras inversum*			
				Speetoniceras inversum				
		Subsaynella sayni		*Crioceratites duvali*	*Crioceratites duvali*			
	lower	*Lyticoceras nodosoplicatum*		*Lyticoceras nodosoplicatum*	*Lyticoceras nodosoplicatum*	*brachiopods are missing*		
		Crioceratites loryi	*Olcostephanus (J.) jeannoti*	*Crioceratites loryi*	? *Crioceratites loryi*	*Musculina sanctaecrucis*	*Symphythyris neocomiensis*	*Cyclothyris corallina neocomiensis* (in reef and shallow facies)
			Crioceratites loryi				*Musculina globus*	
		Acanthodiscus radiatus		*Leopoldia leopoldina*	*Leopoldia desmoceroides*			
Valanginian	upper	*Teschenites callidiscus*		*Eleniceras tauricum*	*Eleniceras tauricum*	?		
				Teschenites callidiscus	*Teschenites callidiscus*			
		Himantoceras trinodosum	*Criosarasinella furcillata*	*Himantoceras trinodosum*	*Himantoceras trinodosum*			
			Olcostephanus (O.) nicklesi					
		Saynoceras verrucosum	*Vahrleideites peregrinus*	*Saynoceras verrucosum*	*Neohoploceras submartini*	*Loriolithyris metaitensis*		
			Karakasch. pronecostatum					
			Saynoceras verrucosum					
	lower	*Busnardoites campylotoxus*		*"Busnardoites" campylotoxus*	*Campylotoxia campylotoxa*			
		Thurmanniceras pertransiens		*Thurmanniceras pertransiens*	*Thurmanniceras pertransiens*			
		Thurmanniceras otopeta		*Kilianella otopeta*	*Kilianella otopeta*	*Loriolithyris carteroniana*		

Fig. 1. Distribution of ammonites and brachiopods in the Valanginian section of the Rezannaya Mountain, localities shown in the inset map by asterisks are (1) Rezannaya and Belaya mountains; (2) Sel'bukhra Mountain; (3) Pervomaiskii quarry (symbols as in Fig. 2).

which are represented by oolitic pudding limestone covered by limonite crust and cemented by younger rocks of a close composition.

In the lowermost 1.5 m of the member, sediments yield ammonites *Kilianella otopeta* (Thieuloy, 1979), *K.* cf. *pexiptycha* (Uhlig), *K. roubaudiana* (d'Orbigny), *Clavithurmannia* aff. *foraticostata* Thieuloy, 1979, *Thurmanniceras* sp., *Bodrakiceras* ex gr. *inostranzewi* (Karakash, 1907), and *B.* sp., which are characteristic of the *Kilianella otopeta* Zone and associate with brachiopods *Praelongithyris credneri* (Weerth), *Loriolithyris valdensis* (Loriol) (Plate, fig. 3), and *L. carteroniana* (d'Orbigny) (Plate, fig. 2).

Ammonites *Thurmanniceras pertransiens* (Sayn), *T. salientinum* (Sayn), *T.* ex gr. *thurmanni* (Pictet et Campiche), *T. valdrumensis* (Sayn), *Bodrakiceras inostranzewi* (Karakash, 1907), and *B. moutonianum* (d'Orbigny) are relatively scarce upward the section. In addition to three brachiopod species found in the *Kilianella otopeta* Zone, *Tropeothyris karakaschi* Smirnova, 1972 (Plate, fig. 1) and *Loriolithyris metaitensis* Middlemiss were encountered at the same level. The described interval corresponds to the *Thurmanniceras pertransiens* Zone.

The eroded HG at the top of the member yields ammonites *Hypophylloceras* cf. *serum* (Oppel), *Eulytoceras phestum* (Matheron), *Thurmanniceras* sp., *Campylotoxia campylotoxa* (Uhlig), *C. campylotoxa densocostata* Baraboshkin et I. Michailova, 2000, *C. evoluta* Baraboshkin et I. Michailova, 2000, *Bodrakiceras moutonianum* (d'Orbigny), *B. inostranzewi* (Karakash, 1907), *B. constrictum* Baraboshkin et I. Michailova, 2000, *B.* sp., *Karakaschiceras biassalense* (Karakash), *Luppovella (Neoluppovella) spinosa* Baraboshkin et I. Michailova, 2000, and *Varlheideites peregrinus* Rawson et Kemper. The assemblage is characteristic of the lower Valanginian *Campylotoxia campylotoxa* Zone.

The assemblage of younger ammonites collected from cementing deposits includes *Busnardoites desori* (Pictet et Campiche), *Neohoploceras submartini* (Mallada), *N. karakaschi* (Uhlig), *N. recticostatum* Baraboshkin et I. Michailova, 1994, *Luppovella (Planibulliceras) kachensis* Baraboshkin et I. Michailova, 1994, *L. (Luppovella) baumbergeri* (Spath), *Rodighieroites*(?) sp., *Neolissoceras subgrasianum* (Drushchits, 1960), and *Valanginites nucleus* (Roemer). Previously the assemblage was assigned to the upper Valanginian *Saynoceras verrucosum* Zone (Baraboshkin 1997a; Baraboshkin and Yanin, 1997; Baraboshkin and Mikhailova, 2000). Now, we should mention, however, the following. (1) *Saynoceras verrucosum* (d'Orbigny) has not been found in the Crimea. (2) The range of *Neohoploceras submartini* (Mallada) is almost identical to that of *Saynoceras verrucosum* (d'Orbigny) in France (Busnardo and Thieuloy, 1979; Reboulet, 1995) and Morocco (collection of Wippich, Bochum, Germany), and close to the last species range in Spain (Company, 1987). (3) *Neohoploceras* is a preferable candidate for the index taxon, since it is in phylogenetic connection with Neocomitinae representing the orthostratigraphic group for all other Valanginian zones. (4) *N. submartini* is of a wider geographic range than *S. verrucosum*. Accordingly, it is reasonable to recognize the new *submartini* Zone in sections of the Crimean Mountains. In the Crimea, the index species of the zone is encountered in the condensed horizon, and hence its exact range is unknown.

Both the *Campylotoxia campylotoxa* and *Neohoploceras submartini* zones are characterized by similar brachiopod assemblages of *Tropeothyris karakaschi* Smirnova, 1972, *Loriolithyris metaitensis* Middlemiss (Plate, fig. 4), and *Sellithyris sella lindensis* Middlemiss 1976. Member *I* ranges in thickness from 0 to 4 m.

(II) The interbedded dense carbonate and loose oolitic sandstones bearing scarce quartz pebble and rare ammonites *Rodigheiroites?* sp., *"Neocomites"* sp., *Neolissoceras subgrasianum* (Drushchits, 1960), and *Spiticeras* sp. The associated fossils are brachiopods *Praelongithyris credneri* (Weerth) (Plate, fig. 5), *Tropeothyris karakaschi* Smirnova, 1972, *Loriolithyris valdensis* (Loriol), *L. metaitensis* Middlemiss, and other faunal remains. Based on stratigraphic position and fauna, we conventionally refer this unit to the *submartini* Zone. The thickness is 4.4–4.5 m.

Members I and II decrease in thickness down to 15–20-cm and constitute a condensed horizon in the east of the Nauchnyi Village, where the rocks contain scarce limonite gravel, quartz pebble, and eroded "pudding" fragments of the Verkhorech'e type, which are covered by limonite crust and yield abundant brachiopods *Sellithyris sella lindensis* Middlemiss 1976. The thickness of the members increases northwards up to 5–6 m, and their facies affinity sharply changes there.

(III) Rhythmically alternating beds of gray calcareous sandstone (0.2–0.3 m thick) and silty to pure bioturbated clays (0.3–1.2 m thick), which are observable on a gentle turf-covered slope. The sandstone beds enclose marcasite and phosphorite concretions and plant detritus. In addition to index species, the zone yields here ammonites *Eulytoceras* cf. *borissiaki* Kuljinskaya-Voronets, *E.* cf. *phestum* (Matheron), *Lytoceras* sp., *Himantoceras* cf. *lessinianum* Faraoni, *Karakaschiceras* sp., *Teschenites* sp., *Neolissoceras grasianum* (d'Orbigny), *N. subgrasianum* (Drushchits, 1960), *Phyllopachyceras* aff. *stuckenbergi* (Karakash, 1907), *P.* sp., and others. Brachiopods are represented by the same species as in Member II. *Sellithyris sella lindensis* Middlemiss 1976, *Praelongithyris credneri* (Weerth), and *Loriolithyris metaitensis* Middlemiss were found in the lower member part in association with *Tropeothyris karakaschi* Smirnova, 1972, and *Loriolithyris valdensis* (Loriol), which occur also slightly higher in the section. The member is 5.9–6.0 m thick.

(IV) The member of interbedded dense calcareous sandstones (0.2–0.3 m thick) and gray loose bioturbated clayey sandstones (1–2 m thick), which bear dense carbonate and marcasite concretions, phosphorite, and large wood fragments. Abundance of phosphatic nodules decreases upward the section. The lower part of the member contains ammonites *Dicostella* cf. *tuberculata* (Roman), *Teschenites* sp. (=*Teschenites neocomiensiformis* (Hohenegger in Uhlig) sensu Thieuloy), *T.* aff. *pachydicranus* Thieuloy, *Oostrella* cf. *cultrata* (d'Orbigny), *Eulytoceras* sp., *Protetragonites* cf. *quadrisulcatum tauricus* Kuljinskaya-Voronets, *Neolissoceras subgrasianum* (Drushchits, 1960), *Hypophylloceras tethys* (d'Orbigny), and *Phyllopachyceras?* sp. in association with brachiopods *Tropeothyris* sp. and other fauna. According to the ammonite assemblage, the member is attributed to the upper Valanginian *Teschenites callidiscus* Zone. Its total thickness is 16.22 m.

(V) Alternating brownish gray, slightly clayey (0.5–2 m) and dense calcareous (0.1–0.3 m) sandstone beds bearing weathered marcasite concretions. A submember of almost pure quartz sandstone is distinguishable at the base. In the section under description, the member yields scarce ammonites, however their representative assemblage was collected in the southern margin of the Nauchnyi Village. The assemblage includes *Eulytoceras konushobaense* Kuljinskaya-Voronets, *Eleniceras tauricum* (Eichwald, 1868), *E. stevrecensis* Breskovski, *E. transsylvanicum* (Jekelius), *E. nikolovi* Breskovski, *E. koeneni* (Karakash, 1907), and *Neolissoceras grasianum* (d'Orbigny). Brachiopods *Tropeothyris* sp. and others were also found. The assemblage is characteristic of the *Eleniceras tauricum* Zone, the terminal one in the Valanginian succession. The total thickness of the member is 17.5 m. In the area between the Sel'bukhra and Nauchnyi villages, the zone section is condensed, and *Eleniceras* forms are encountered here together with the lower Hauterivian *Leopoldia desmoceroides* (Karakash, 1907), *Crioceratites?* sp., and other taxa.

The Valanginian sediments of a different structure are exposed near the Prokhladnoe Village (Patil', Sheludivaya, and Dlinnaya mountains) and on the right bank of the Bodrak River (Bol'shoi and Malyi Kermen mountains). The most representative section exposed on the northwestern slope of the Dlinnaya Mountain consists of the following members (from the base upward).

(I) The member of rhythmically alternating cross-bedded compact to loose sandstones calcareous or dolomitized, overlying with a distinct unconformity the Tavrian rocks and bearing dispersed pebble at the base. The sediments yield scarce ammonites *Kilianella otopeta* (Thieuloy, 1979), *K.* cf. *pexiptycha* (Uhlig), *K.* cf. *roubaudiana* (d'Orbigny), *Thurmanniceras gueymardi* (Sayn), *T.* spp., *Distoloceras* sp., *Neolissoceras grasianum* (d'Orbigny), *Luppovella (Planibulliceras) lambertiformis* Baraboshkin et I.Michailova, 1994, brachiopods *Dictyothyris manguschensis* Smirnova, and other fauna. The total thickness is 1.5–2.5 m.

(II) Rhythmic alternation of massive and loose calcareous sandstones, sometimes oolitic and cross-bedded; the sediments are looser than the underlying member and completely replaced by the lower Hauterivian deposits in the southern part of the mountain. The member yields *Thurmanniceras* cf. *pertransiens* (Sayn), *T. salientinum* (Sayn), *T. gueymardi* (Sayn), *T.* sp., *Bodrakiceras inostranzewi* (Karakash, 1907), *B. moutonianum* (d'Orbigny), *B.* cf. *trezanensis* (Lory in Sayn), and *B.* sp. The thickness ranges from 0 to 6 m.

Brachiopods *Loriolithyris valdensis* (Loriol) and *Terebrataliopsis compacta* sp. nov. are encountered in analogous lower Valanginian calcareous sandstones of the Sheludivaya Mountain. The latter species together with *Terebrataliopsis semicircularis* sp. nov., was also found in the section of the Patil' Mountain (Plate, figs. 13, 14).

The Hauterivian sediments are best exposed in the gully on the southern slope of the Belaya Mountain (Fig. 2) above western outskirts of the Verkhorech'e Village. The following sequentially numbered members are distinguished here (from the base upward).

(VI) Alternating beds of gray loose, bioturbated, clayey to gravelly sandstone (0.2–1 m thick) and compact, grayish brown, calcareous-detrital gravelly sandstones (0.1–0.3 m) bearing marcasite concretions. The member unconformably overlies rocks of the upper Valanginian *Eleniceras tauricum* Zone. Pebbles and black phosphorite concretions, commonly the internal molds of echinoids, occur within the 1.55-m-thick interval above the member base. The soft (SG) and hard ground levels separate dense and loose sandstone beds. The gravel abundance decreases upward the section. The member yields ammonites *Lytoceras?* sp., *Leopoldia desmoceroides* (Karakash, 1907), *Neolissoceras grasianum* (d'Orbigny), brachiopods *Musculina nuciformis* sp. nov., and other fauna. Ammonites *Breistrofferella castellanensis* (d'Orbigny) characteristic of the Hauterivian base in many regions, are encountered in sandstones exposed along the road from the Nauchnyi to Verkhorech'e Village. The same ammonites were found in association with *Leopoldia* cf. *leopoldina* (d'Orbigny) and *L. desmoceroides* (Karakash, 1907) in the condensed horizon sampled in upper reaches of the Koyas-Dzhilga Gully. Ammonites of *Leopoldia* and *Breistrofferella* genera suggest the member correlation with the lower Hauterivian *Acanthodiscus radiatus* Zone of the standard Mediterranean zonation (Hoedemaeker and Rawson, 2000). Originally this interval was distinguished as the *Leopoldia leopoldina* Zone (Baraboshkin, 1997a, 1997b), but single specimen of this index species was identified in the open nomenclature and found in the talus. Shells of *Leopoldia desmoceroides* (Karakash, 1907) are much more numerous in the Kacha, Bodrak, and Al'ma river basins. The species

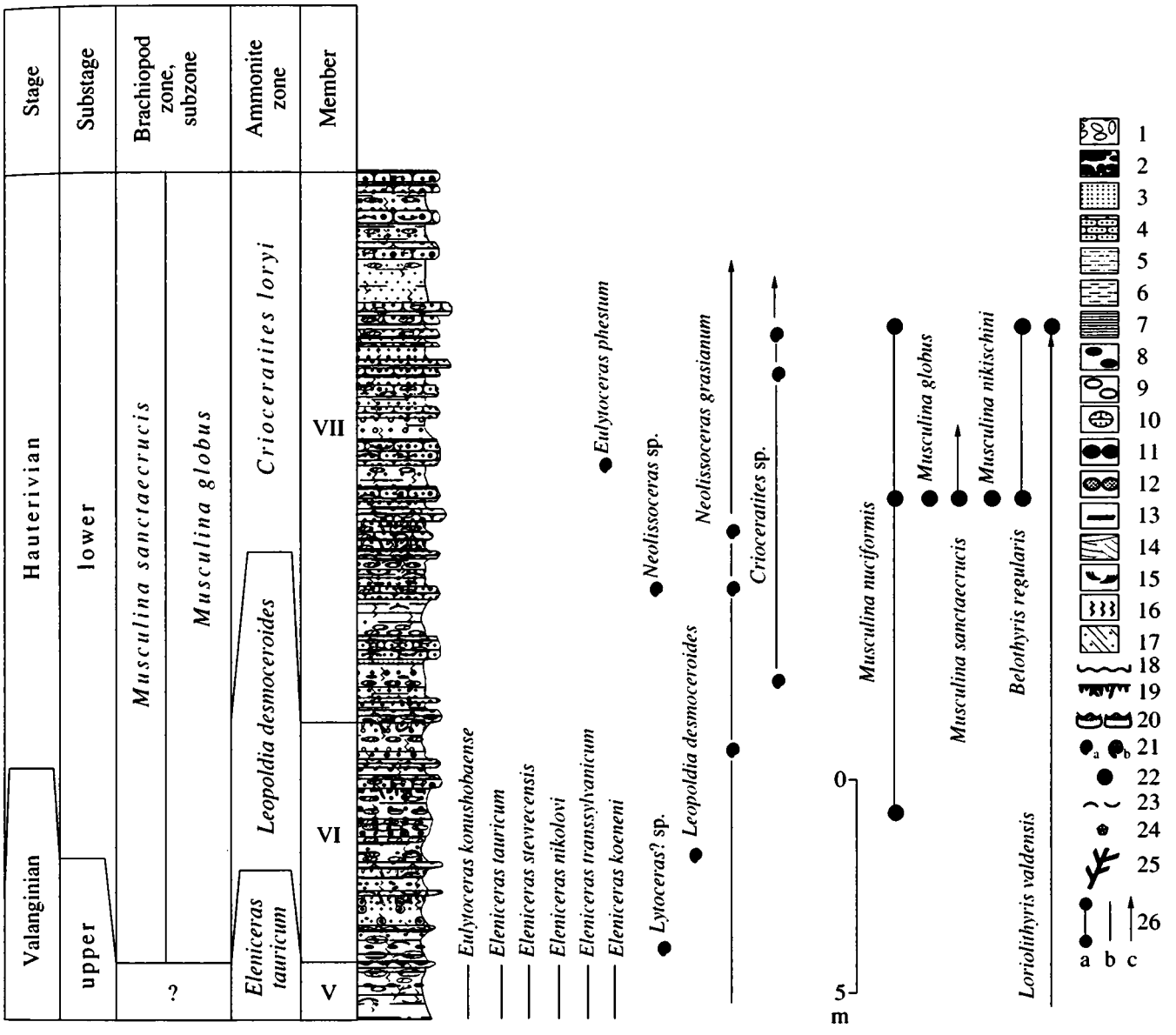

Fig. 2. Distribution of ammonites and brachiopods in the lower Hauterivian section of the Belaya Mountain: (1) conglomerate; (2) phosphorite conglomerate; (3) sand, loose sandstone; (4) compact sandstone; (5) silt, siltstone; (6) argillite, clayey rock; (7) clay; (8) phosphorite pebbles; (9) diverse pebbles; (10) sandy-carbonate concretions; (11) Fe-sulphide concretions; (12) Fe-carbonate concretions; (13) ferruginous rock; (14) diagonal bedding; (15) cross-bedding; (16) bioturbation; (17) terrigenous flysch of the Tavrian Group; (18) erosional surfaces; (19) soft or hard ground surface; (20) eroded hardground surface; (21) ammonites found *in situ* (a) or redeposited (b); (22) brachiopods; (23) shelly detritus; (24) crinoids; (25) plant detritus; (26) faunal remains allied with beds (a) or of unknown alliance (b) and intervals with possible faunal remains (c); arrows indicate the species occurrence upward or downward the section.

was also found in the northern Caucasus (Kvantaliani *et al.*, 1990). Therefore, we suggest to replace the *leopoldina* index species with *desmoceroides*. The replacement seems reasonable, since the established zone range is inconsistent with that traditionally accepted. Previously the Valanginian and a part of the lower Hauterivian was erroneously referred to as the *Leopoldia leopoldina* Zone. However, "*Leopoldia leopoldi* (d'Orbigny)" illustrated by Eristavi (1957, Plate IV, figs. 3, 4) and Drushchits (Drushchits and Kudryavtsev, 1960, Plate XXVIII, fig. 3) actually represents the lower Valanginian species *Bodrakiceras inostranzewi* (Karakash, 1907).

The total thickness of the member is 5.6 to 6 m. It sharply decreases northward. In upper reaches of the Koyas-Dzhilga Gully and near the Nauchnyi Village, the member is condensed, and ammonites of the zone were encountered in the eroded HG fragments together with species of the overlying *nodosoplicatum* Zone.

(VII) Rhythmically alternating beds, which are composed of dark gray, loose, clayey, bioturbated, detrital sandstone (0.2–0.5 m thick) bearing quartz and limonite gravel, and of compact calcareous, grayish brown gravelly sandstone (0.1–0.6 m thick) with marcasite concretions. The thickness of rhythms decreases upward the section from 1.2–1.5 to 0.5–0.6 m. They are commonly separated by the HG surfaces. At the base and 2.7 m above it, black phosphorite pebbles and small (up to 1 cm) grains are seen in sediments. At the level of 1.45 m above the base, the sandy rhythm 1.25 m thick is almost lacking gravely fraction. The member contains ammonites *Crioceratites* sp., *Neolissoceras grasianum* (d'Orbigny), *N.* sp., *Lytoceras?* sp., and *Eulytoceras phestum* (Matheron). Brachiopods are represented by six species, among them *Musculina nuciformis* sp. nov. is known from Members VI and VII, whereas *Belothyris regularis* Smirnova and the scarcer *Musculina nikischini* sp. nov., *M. globus* (Pictet) (Plate, fig. 8), *M. sanctaecrucis* (Gatzigras), and *Loriolithyris valdensis* (Loriol) occur in the middle of the member. The total thickness is 13 to 14 m. It decreases northwards, where sediments are condensed.

(VIII) The member similar to the underlying one is less clayey and grayish brown in coloration. It almost lacks gravel, except for the uppermost part. Phosphorites have not been found, marcasite concretions are scarce, and loose silty and sandy beds are of a greater thickness (up to 0.5–1.25 m). Additionally, the erosion surfaces are common between dense and loose layers. The top is eroded. Precisely this part of the Hauterivian section is responsible for characteristic stepwise topography on the southern slope of the Belaya Mountain. The corresponding rocks armor also the top of the Rezannaya Mountain. The member yields ammonites *Neolissoceras grasianum* (d'Orbigny, 1840) and brachiopods *Belothyris regularis* Smirnova and *Musculina globus* (Pictet). The total thickness is 12–13 m.

Members VII and VIII are conventionally attributed to the *Crioceratites loryi* Zone based on abundant fragments of *Crioceratites* sp., though the index species itself has not been found.

(IX) Dense clayey to gravelly sandstone beds (0.1–0.5 m thick) bearing marcasite concretions and phosphorites in the upper part are intercalated with dark gray to greenish, loose clayey sandstone and clay beds (0.15–1.65 m thick). The limonite and hydrogoethite ooliths and several ocherous HG surfaces occur near the top. The member contains ammonites *Crioceratites* sp., *Lyticoceras cryptoceras* (d'Orbigny), *L.* sp., *Phyllopachyceras stuckenbergi* (Karakash, 1907), *Spitidiscus rotula inflatum* (Kilian), *Abrytusites* sp., *Hamulina* sp., *Eulytoceras* sp., and other fauna but not brachiopods. The total thickness is 7.5–8 m.

(X) Gray bioturbated silty clays, sandy in the lower three meters, bearing plant remains and rare *Hibolites* sp. The top of the member is eroded. The thickness is 9.5 m.

Based on occurrence of *Lyticoceras cryptoceras* (d'Orbigny) and stratigraphic position, members IX and X are referred to as the *Lyticoceras nodosoplicatum* Zone distinguished in France, England, and other regions. These members overlying different rocks are traceable northwards up to the Kizil-Chigir Mountain, where their facies affinity changes.

(XI) Sedimentary rhythms composed of the following rock types: (1) grayish green bioturbated clayey silts (0.05–0.23 m thick), which bear plant detritus, small phosphorite pebble, rounded rostra of *Hibolites* sp., limonite gravel and grade upward into almost pure clay; (2) brown bioturbated oolitic limestone (0.05–0.25 m thick), the erosion top of which is ocherous and represents the HG. Ammonites *Crioceratites duvali* (Leveille) were found in the upper limestone bed by Kakabadze (1981). We encountered *Crioceratites* cf. *tenuicostatus* Thomel, 1964, *C.* aff. *duvali* (Leveille), and *C.* sp. at that level. These forms occur also higher, at the base of the "Cephalopod Limestone" and in the fragments of the eroded mature HG covered by a thick limonite crust, when sediments are condensed. The thickness of the member is 1.7 m.

The member corresponds to the *Crioceratites duvali* Zone according to occurrence of the index species. In the Crimea, the zone was recognized as an equivalent of the entire upper Hauterivian (Eristavi, 1957), being renamed afterward into the upper Hauterivian *Crioceratites duvali–Speetoniceras versicolor* Zone (Drushchits and Kudryavtsev, 1960). After the work by Thomel (1964) who studied of the *duvali* distribution throughout the Hauterivian of southeastern France, the unit was not recommended for use (Kakabadze, 1981). The latest data on this species distribution in the Hauterivian sediments of the stratotype area indicate, in distinction from the *nolani* form, that it appears concurrently with *Subsaynella sayni* (Paquier) and disappears slightly below the base of the overlying *balearis* Zone (Autran, 1989; Ropolo, 1995). Thus, the biozone in question is almost identical to the *sayni* Zone. According to data of Eristavi (1957), Drushchits and Kudryavtsev (1960), Kakabadze (1981), and our own, the species occurs immediately below the base of the "Cephalopod Limestone" and, probably, at the base of the latter in the redeposited state along with *Speetoniceras*. Therefore, it is possible to distinguish the individual *duvali* Zone correlative with the *sayni* Zone of the stratotype. The substage boundary (Table 1) should be placed at the zone base but not at the top as suggested before (Baraboshkin, 1997a, b).

Members I–IX of the Verkhorech'e section decrease in thickness north- and northeastward being considerably condensed in the basal portion of Member X. Such condensed section is known in upper reaches of a confluent of the Koyas-Dzhilga Gully crossing the Observatoriya Plateau between the Nauchnyi Village and Sel'bukhra Mountain (Fig. 3). Here, the eroded and ocherous surface of the upper Valanginian *Eleniceras*

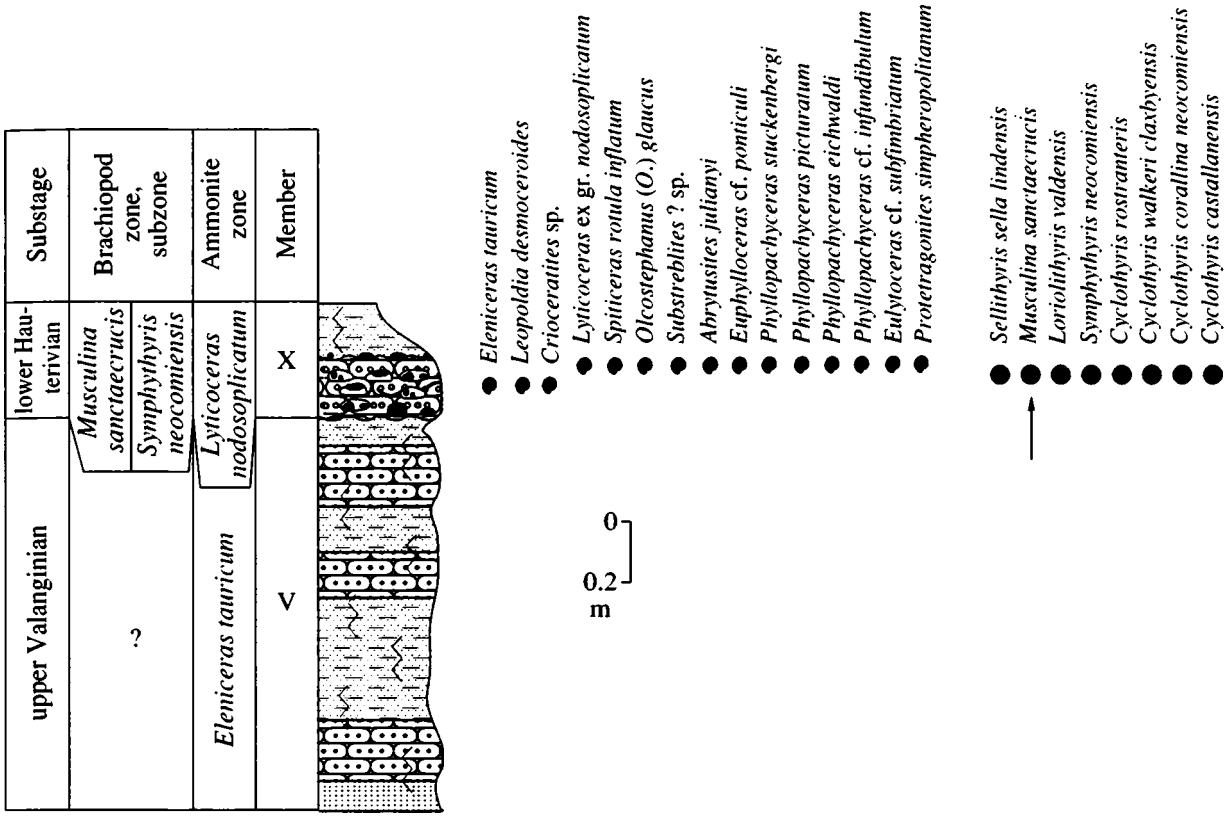

Fig. 3. Distribution of ammonites and brachiopods in the lower Hauterivian section of the Sel'bukhra Mountain (symbols as in Fig. 2).

tauricum Zone is overlain by the bed of large (10–12 cm) phosphatic blocks of corroded sandstone that represent a mature HG. The blocks resembling rounded boulders are not redeposited, however, being partially dissolved and overgrown with serpulids and oysters. They are cemented into a single "plate" by the calcareous sandy matrix bearing well-rounded quartz gravel, later generation of phosphorite grains, shelly detritus, and numerous mollusks, which either redeposited or buried *in situ*. The "plate" is 0.15–0.2 m thick. Among the redeposited ammonites, the upper Valanginian *Eleniceras* were found in association with lower Hauterivian *Leopoldia* and *Crioceratites*. The gravelstone yields ammonites *Crioceratites* sp., *Lyticoceras* ex gr. *nodosoplicatum* (Kilian et Reboul), *Neolissoceras grasianum* (d'Orbigny), *Hypophylloceras* cf. *ponticuli* (Rousseau), *Phyllopachyceras stuckenbergi* (Karakash, 1907), *P. picturatum* (d'Orbigny), *P. eichwaldi* (Karakash), *P.* cf. *infundibulum* (d'Orbigny), *Spitidiscus rotula inflatum* (Kilian), *Olcostephanus (O.) glaucus* Spath, *O.* (*O.*) sp., *Substreblites?* sp., *Abrytusites julianyi* (Honnorat-Bastide), *A.* sp., *Eulytoceras* cf. *subfimbriatum* (d'Orbigny), *E.* sp., *Protetragonites simpheropolitanum* (Kuljinskaya-Voronets), and *P.* sp. These coexisted with nautiloids, belemnites, bivalves, gastropods, scaphopods, echinoderms, shark teeth, burrowed wood remains, and brachiopods *Symphythyris neocomiensis* (d'Orbigny) (Plate, fig. 11), *Musculina sanctaecrucis* (Gatzigras) (Plate, fig. 7), *Belothyris regularis* Smirnova (Plate, fig. 10), *Loriolithyris valdensis* (Loriol), *Cyclothyris rostriformis* (Roemer), *C. walkeri claxbyensis* (Owen et Thurrell) (Plate, fig. 16), *C. castellanensis* (Jacob et Fallott) (Plate, fig. 17), and *C. corallina neocomiensis* (Jacob et Fallott) (Plate, fig. 18). *Sellithyris sella lindensis* Middlemiss 1976 may be a redeposited component of this assemblage. The ammonite assemblage is analogous to that of the West European *Lyticoceras nodosoplicatum* Zone in the stratotype and elsewhere (Busnardo *et al.*, 1989).

The facies composition of Member X rapidly changes in the northeastern direction, and in the Prisyazhnaya Mountain section it encloses the lenticular beds of coarse- to medium-grained detrital carbonate sandstones yielding large bivalve remains, solitary corals buried *in situ* replaced by coarse crystalline calcite, and well-rounded redeposited fragments of colonial corals. In the area of the Prisyazhnaya, Dlinnaya, Patil', and Bol'shoi Kermen mountains, rocks of this facies unconformably overlie the lower Valanginian sediments. These rocks contain brachiopods *Cyclothyris corallina neocomiensis* (Jacob et Fallott) and *C. irregularis* (Pictet), on the Patil' Plateau, and *Cyclothyris rostriformis* (Roemer) in the Dlinnaya Mountain

section. The apparent thickness of Member X in this area is 2 to 3 m.

On the right bank of the Bodrak River, a thin coral reef and its apron facies correspond to sandy rocks of the *desmoceriodes–loryi* interval in the Kacha section. The reef is best exposed in the eastern part of the Pervomaiskii quarry, where it rests above the strongly eroded and weathered gabbro-diorite, being composed of tightly arranged reniform and massive coral colonies up to 1 m thick, by solitary corals, and sponges. The colonies are in the lifetime position with respect to enclosing sediments, more rarely they are inversed. Corals are strongly recrystallized, slightly rounded sometimes, burrowed, and incrusted by bivalves, bryozoans, and serpulids. The space between colonies is filled with a carbonate matrix composed of coral sand, gravel, and shelly detritus. Accumulations of brachiopods, echinoids, and other fauna are common in large cavities. Brachiopods are represented by *Loriolithyris valdensis* (Loriol), *Cyclothyris corallina neocomiensis* (Jacob et Fallott), *C. ardescica* (Jacob et Fallott), *C. irregularis* (Pictet), *C. desori* (Pictet), *C.* ex gr. *renauxiana* (d'Orbigny), *C. castellanensis* (Jacob et Fallott), and *C. rostriformis* (Roemer). The reef thickness varies from 0.6 to 3 m within the quarry and in the area of the Kizil-Chigir Mountain. Above the contact of the Pervomaiskii intrusion with rocks of the Eskiordan Group exposed in the northern wall of the quarry, the coral reef grades into the apron facies extending west- and northwestward. The facies is confined to distribution area of clayey rocks of the Eskiordan and Tavrian groups, whereas the reef rests on the hard substratum of the intrusion and volcanogenic rocks of the Karadag Group, a continuous belt of which extends from the Trudolyubovka Village to Simferopol. The apron facies overlies with a hiatus the lower Valanginian sandy dolomitized limestones. Its thickness is 3–4 m in the Pervomaiskii quarry and up to 10–12 m on the Malyi Kermen Mountain. The reefal and apron facies are defined, based on their stratigraphic position as the Early Hauterivian in age (Baraboshkin, 1997a, 1997b; 2001).

The top of the coral reef is crowned by the HG in the Pervomaiskii quarry. It is eroded, burrowed, and limonitized. In the member, which overlies the reef, brown compact cross-bedded calcareous sandstones up to 0.6 m thick bear abundant solitary corals *Cunnolites intumescens* (Trautschold), small bivalves, sponges, and other fauna, whereas brown loose, slightly bioturbated sandstones and sands 0.3 to 0.5 m thick show presence of *Panopea neocomiensis* (Leymerie) buried in the lifetime position. Brachiopods *Cyclothyris corallina neocomiensis* (Jacob et Fallott), *C. castellanensis* (Jacob et Fallott), and *C. desori* (Pictet) were encountered in that member as well. This brachiopod assemblage is similar in composition to the later (not redeposited) brachiopod generation from the condensed bed on the southern slope of the Sel'bukhra Mountain. The species in common are *C. castellanensis* (Jacob et Fallott), *C. corallina neocomiensis* (Jacob et Fallott), *C. rostriformis* (Roemer), and *Loriolithyris valdensis* (Loriol). The apparent thickness of the member is 3 to 4 m near the Pervomaiskii quarry and decreases northward to a complete wedging out near the Kizil-Chigir Mountain. The sandstones yield very rare ammonites *Spitidiscus rotula inflatum* (Kilian) and *Lyticoceras* sp. characteristic of the *Lyticoceras nodosoplicatum* Zone.

ANALYSIS OF BRACHIOPOD ASSEMBLAGES

The generalized stratigraphic distribution of the studied brachiopods is shown in Table 2. The forms encountered in shallow- and relatively deep-water facies are regarded separately (Baraboshkin, 1997b), since their distribution reveals certain regularities.

According to our data, brachiopods found in three lower members of the Valanginian section on the right bank of the Kacha River valley near the Verkhorech'e Village are represented by six species.

Loriolithyris carteroniana (d'Orbigny) was encountered in the lower and middle parts of Member I corresponding to the lower Valanginian *Kilianella otopeta* and *Thurmanniceras pertransiens* zones. We propose to distinguish this interval of the section as the *Loriolithyris carteroniana* Zone. This species is abundant in many areas: in the Valanginian of the Swiss Jura, i.e., in limestones known as the Calcaire Roux; in the Valanginian–Hauterivian of the southeastern Parisian basin, and in the Hauterivian of southwestern France, the Aude Province. It was also described from the condensed upper Valanginian–lower Hauterivian beds of northern Germany (Grenzlerburg, Salzgitter). In Russia, the species is known from the northwestern Caucasus, where it was found in the Belaya River sequence of interbedded sandstone and clayey siltstone of the lower and lowermost upper Hauterivian and at the base of the lower (?) Hauterivian limestone bed bearing sandstone pebble, the Kuban' River exposures (Drushchits and Mikhailova, 1966).

Species *Loriolithyris valdensis* (Loriol), *L. metaitensis* Middlemiss, *Praelongithyris credneri* (Weerth), *Tropeothyris karakaschi* Smirnova, 1972, and *Sellithyris sella lindensis* Middlemiss, 1976, were found in three lower members, which correspond to the uppermost *pertransiens* Zone of the lower Valanginian, to the *Neohoploceras submartini* Zone, and to the lower part of the upper Valanginian *Himantoceras trinodosum* Zone. We distinguish this interval as the *Loriolithyris metaitensis* Zone.

Loriolithyris metaitensis Middlemiss is known from the Hauterivian–Barremian sediments of southwestern Morocco. *Praelongithyris credneri* (Weerth) was encountered in the upper Valanginian–lower Hauterivian sediments of Germany (Grenzlerburg) and northwestern England (Claxby Ironstone), and also in the Hauterivian part of the Speeton Clay, northern England. *Sellithyris sella lindensis* Middlemiss, 1976, was described from the upper Valanginian–lower Hauteriv-

Table 2. Stratigraphic range of the studied brachiopods: dashed line designates species characteristic of relatively shallow facies and crosses, deep-water taxa.

Stage	Valanginian							Hauterivian		
Substage	lower			upper				lower		
Ammonite zone / Brachiopod species	*otopeta*	*pertransiens*	*campylotoxa*	*submartini*	*trinodosum*	*callidiscus*	*tauricum*	*desmoceroides*	*loryi*	*nodosoplicatum*
Symphythyris neocomiensis										---
Cyclothyris walkeri claxbyensis										---
Cyclothyris castellanensis								----------		
Cyclothyris corallina neocomiensis								----------		
Cyclothyris desori								----------		
Cyclothyris irregularis								----------		
Cyclothyris rostriformis								----------		
Cyclothyris ex gr. renauxiana								----------		
Cyclothyris ardescica								----------		
Dictyothyris manguschensis	----------									
Terebrataliopsis compacta	----------									
Terebrataliopsis semicircularis	----------									
Musculina sanctaecrucis									+++++	
Belothyris regularis									++	
Musculina globus									++	
Musculina nikischini									++	
Musculina nuciformis								+++++		+++
Sellithyris sella lindensis				++++++++						
Tropeothyris karakaschi		++++++++++++++++								
Loriolithyris metaitensis		++++++++++++++++								
Praelongithyris credneri	++++++++++++++++++									
Loriolithyris carteroniana	+++									
Loriolithyris valdensis	++++++++++++++++++								++++	----

ian condensed beds of northern Germany and northwestern England (Middlemiss, 1976). The other species, except for *Loriolithyris valdensis* (Loriol), are characteristic of the Valanginian sediments of the southwestern Crimea. The latter form occurs in both the Valanginian and lower Hauterivian sediments of the Verkhorech'e section.

The lower Valanginian facies of shallower settings are exposed near the Prokhladnoe Village and on the right flank of the Bodrak River valley. They yield a less diverse brachiopod assemblage of *Terebrataliopsis semicircularis* sp. nov., *T. compacta* sp. nov. and endemic *Dictyothyris manguschensis* Smirnova.

The lower Hauterivian assemblage of relatively deep-water brachiopods from the Verkhorech'e section is composed of six species, two among them are first described: *Musculina nuciformis* sp. nov. and *M. nikischini* sp. nov. The four other species described previously are *Musculina sanctaecrucis* (Gatzigras), *M. globus* (Pictet), *Loriolithyris valdensis* (Loriol), and *Belothyris regularis* Smirnova. *Musculina nuciformis* sp. nov. was found in the upper part of Member VI corresponding to the *Leopoldia desmoceroides* Zone, and in Member VII equivalent to the *Crioceratites loryi* Zone. *Musculina sanctaecrucis* (Gatzigras) was encountered in the middle of Member VII correspond-

ing to the lower part of the *Crioceratites loryi* Zone. *M. sanctaecrucis* is characteristic of the lower Hauterivian organogenic limestones of the Western Swiss (Landeron, Sainte-Croix, Auberson, and Morteau sites) and French Jura (Sanseau coral limestones), and of the biogenic limestones in southeastern France (Basses Alps, Var, Castellane). It was also found in condensed beds attributed to the upper Valanginian–lower Hauterivian in Germany. *Musculina sanctaecrucis* (Gatzigras) is known from the lower Hauterivian marls in the northern Caucasus (Ol'khovka River, Dagestan), and from the Hauterivian of the Transcaucasia and western Turkmenistan. *Musculina globus* (Pictet) occurring at the same level, is known from the upper Hauterivian–lower Barremian of Switzerland. *Loriolithyris valdensis* (Loriol) was found in the upper part of Member VII corresponding to the *Crioceratites loryi* Zone. It is also dispersed throughout the Valanginian sediments in the Verkhorech'e section. *Loriolithyris valdensis* (Loriol) was previously described from the Valanginian of the Swiss Jura (Sainte-Croix, Villers-le-lac, Arzier) and from the lower Hauterivian of the northern Caucasus (Kuma River) and Dagestan. *Belothyris regularis* Smirnova was described from the lower Hauterivian of the northern Caucasus and Georgia.

All the species of the lower Hauterivian brachiopod assemblage are mainly characteristic of the *Crioceratites loryi* Zone. These are *Musculina nuciformis* sp. nov., *M. nikischini* sp. nov., *M. globus* (Pictet), *M. sanctaecrucis* (Gatzigras), and *Belothyris regularis* Smirnova. *Musculina nuciformis* sp. nov. was also encountered in the sediments of the *Leopoldia desmoceroides* Zone. Therefore, we distinguish the *Musculina globus* Subzone in the lower part of the *sanctaecrucis* Zone.

The brachiopod assemblage of eight species, which are mainly known from the Valanginian–lower Hauterivian of Western Europe and lower Hauterivian of the northern Caucasus, was collected from a condensed bed on the southern slope of the Sel'bukhra Mountain. Three species were found separately in the Valanginian or lower Hauterivian near the Verkhorech'e Village. The assemblage includes *Symphythyris neocomiensis* (d'Orbigny) and *Sellithyris sella lindensis* Middlemiss 1976 (redeposited?), *Musculina sanctaecrucis* (Gatzigras), *Loriolithyris valdensis* (Loriol), *Cyclothyris rostriformis* (Roemer), *C. walkeri claxbyensis* (Owen et Thurrell), *C. castellanensis* (Jacob et Fallott), and *C. corallina neocomiensis* (Jacob et Fallott). The last one occurs in the Valanginian of Switzerland. *Musculina sanctaecrucis* (Gatzigras), *Cyclothyris castellanensis* (Jacob et Fallott), and *C. rostriformis* (Roemer) are known only from the Hauterivian sediments. *C. castellanensis* was described from the lower Hauterivian marls and biogenic limestones of southeastern France (Lower Alps, Var, Castellane) and from the Hauterivian strata exposed along the Kuban' River in the northern Caucasus. *Cyclothyris rostriformis* (Roemer) occurs in the lower Hauterivian limestones of the Swiss Jura, Parisian Basin, and northwestern Germany. *Sellithyris sella lindensis* Middlemiss, 1976, was encountered in the Verkhorech'e section in the upper Valanginian sediments corresponding to the *Neohoploceras submartini* and lower part of the *Himantoceras trinodosum* zones. It also occurs in the condensed Claxby Ironstone (upper Valanginian–lower Hauterivian) of northwestern England and in Germany. *Cyclothyris walkeri claxbyensis* (Owen et Thurrell) is known from the upper Valanginian–lower Hauterivian sediments in England and Germany as well. *Symphythyris neocomiensis* (d'Orbigny) was found in the Valanginian marls of Switzerland, in the lower Hauterivian organogenic limestones of the Swiss Alps, in the lower Hauterivian coral limestones of France, in the Hauterivian limestones of the northwestern Caucasus (Belaya River basin), and in the lower (?) Hauterivian of the Kuban' River (Drushchits and Mikhailova, 1966). In the studied sections it is associated only with the lower Hauterivian *nodosoplicatum* Zone, and we propose therefore to consider this interval as the *Symphythyris neocomiensis* Subzone of the brachiopod *Musculina sanctaecrucis* Zone.

In the shallow biohermal limestones of the Pervomaiskii quarry, which correspond to the *desmoceriodes–loryi* zones of the Verkhorech'e section, brachiopods are represented by species widespread in Western Europe and the northern Caucasus. These are the aforementioned *Loriolithyris valdensis* (Loriol), *Cyclothyris corallina neocomiensis* (Jacob et Fallott), *C. rostriformis* (Roemer), and *C. castellanensis* (Jacob et Fallott). *Cyclothyris ardescica* (Jacob et Fallott) is known from the lower Hauterivian of France and Switzerland and from the upper Valanginian–lower Hauterivian of Sardinia. *Cyclothyris irregularis* (Pictet) is characteristic of the lower Hauterivian in the northern Caucasus, of the Hauterivian in Dagestan, and of the Barremian in France and Switzerland. *Cyclothyris desori* (Pictet) was encountered in the Valanginian of France, Switzerland, and southwestern Crimea, and in the Hauterivian of Azerbaijan. The biohermal limestones also yield *Cyclothyris* ex gr. *renauxiana* (d'Orbigny). The most widespread among brachiopods is *Cyclothyris corallina neocomiensis* (Jacob et Fallott) occurring in both the reef limestones and shallow sandy cross-bedded facies. Therefore, we consider the range of this species as characterizing the lower Hauterivian zone in shallow facies, in addition to the *Musculina sanctaecrucis* Zone that characterizes the deeper-water facies.

The upper Valanginian and lower Hauterivian brachiopod assemblages from the studied sections of the southwestern Crimea have no species in common except for *Loriolithyris valdensis* (Loriol) distributed throughout the Valanginian and lower Hauterivian. The refined stratigraphic ranges of the discussed species based on correlation with ammonite zones make them usable as the guide and zonal index taxa for age determination of sediments in southern Russia and Western Europe.

Thus, the studied distribution of brachiopod species allowed us to suggest a new brachiopod biostrati-

graphic zonation correlated with the ammonite standards (Table 1). The zonation is based on the widespread index species also known from many regions of Russia and other countries. Its advantage is a considerable versatility in application to various facies. However, the exact stratigraphic position (in terms of biozonation) of the selected index species is yet unknown in many regions, and therefore a certain diachronism of suggested zonal boundaries can be expected because of this disadvantage.

DESCRIPTION OF THE NEW TAXA

Order Terebratulida Waagen, 1883

Superfamily Terebratuloidea Gray, 1840

Family Sellithyrididae Muir-Wood, 1965

Genus *Musculina* Schuchert et Le Vene, 1929

Musculina nuciformis Smirnova sp. nov.

Plate, fig. 6

Name derived from *nucis* Lat.—nut, and *forma*—shape.

Holotype: 1/100, Museum of Earth Science, Moscow State University, southwestern Crimea, the Kacha River, Verkhorech'e Village; lower Hauterivian.

Material: 25 well-preserved specimens and 27 slightly compressed specimens from the Verkhorech'e section; 46 well-preserved specimens from the foot of the Sel'bukhra Mountain.

Description. Shells are medium-sized, variable in outline from rounded-pentagonal to oval. The anterior commissure is arched. Lateral commissures are curved as a high arc in the anterior half of the shell. Shells are maximum wide in the anterior one-third of their length and most thick in the middle. Lateral margins are moderately convex. Ventral valve is equally or less convex than dorsal one, being uniformly moderately convex in transverse and longitudinal section. The anterior half of the valve is flattened. The anterior margin is monoplicate, tongue-shaped, and dorsally steeply inclined. Lingula is arched or trapeziform in outline. Beak is low, slender, curved, with smoothed beak keels. False area is low, slightly concave. The mesothyridid foramen is large. Apical angle is 67°–83°. Dorsal valve is strongly convex in the beak half. A fold, flattened in the middle and bordered by shallow lateral depressions, is characteristic of the anterior half of shell.

Size parameters in mm and their ratios:

Sample	L	W	Th	W/L	Th/L
Holotype 1/100	16.2	13.0	10.0	0.80	0.61
5/100	15.9	12.7	10.0	0.80	0.64
6/100	15.0	10.7	9.6	0.71	0.64
7/100	13.4	11.0	8.3	0.82	0.62
8/100	13.0	11.0	8.0	0.85	0.61

Comparison. In distinction from *Musculina sanctaecrucis* (Gatzigras), shells are less in size, having more convex valves, monoplicate anterior margin, lower beak, and larger apical angle. Monoplicate anterior margin, arched anterior commissure, sharply curved lateral commissures, disposition of greatest width, and narrow hinge plate differ them from *Musculina globus* (Pictet, 1872).

Distribution: lower Hauterivian of the southwestern Crimea.

Musculina nikischini Smirnova sp. nov.

Plate, fig. 9

Named in honor of A.M. Nikishin.

Holotype: 2/100, Museum of Earth Science, Moscow State University, southwestern Crimea, the Kacha River, Verkhorech'e Village; lower Hauterivian.

Material: 69 well-preserved specimens from the Verkhorech'e section.

Description. Small oval shells with equally strongly convex valves have biplicate anterior margin and gentle W-shaped anterior commissure. Lateral commissures are arched near the anterior margin. Shells are maximum wide one-third away from the anterior margin, being most thick in the middle or slightly closer to beak. Cardinal margin is strongly curved and lateral margins are strongly convex. Ventral valve is extremely convex near beak, uniformly strongly convex in the middle, and moderately convex near the anterior margin. A low smoothed plica is poorly distinguishable near the anterior margin. Beak is high, slender, curved, with poorly developed ridges. False area is barely expressed. The mesothyridid foramen is circular, of a medium size. Apical angle is 65–78°. Dorsal valve is strongly and uniformly convex throughout its surface. Two short low plicae separated by an insignificant sulcus are distinct near the anterior margin. On each side of the plicae, there are two shallow depressions.

Size parameters in mm and their ratios:

Sample	L	W	Th	W/L	Th/L
Holotype 2/100	16.1	13.0	9.8	0.81	0.61
10/100	15.0	10.7	9.2	0.71	0.61
11/100	14.7	10.2	10.7	0.69	0.74
12/100	14.2	10.8	9.5	0.76	0.67
13/100	13.3	10.0	8.1	0.75	0.61
14/100	13.1	10.3	8.7	0.78	0.66

Comparison. In distinction from *Musculina nuciformis* sp. nov., shells, being less wide, have biplicate anterior margin, W-shaped anterior commissure, high beak, convex anterior margin, hamate teeth, rimose dental sockets, and narrow hinge plates. Smaller size, slightly plicated valves, gentle W-shaped anterior com-

Plate

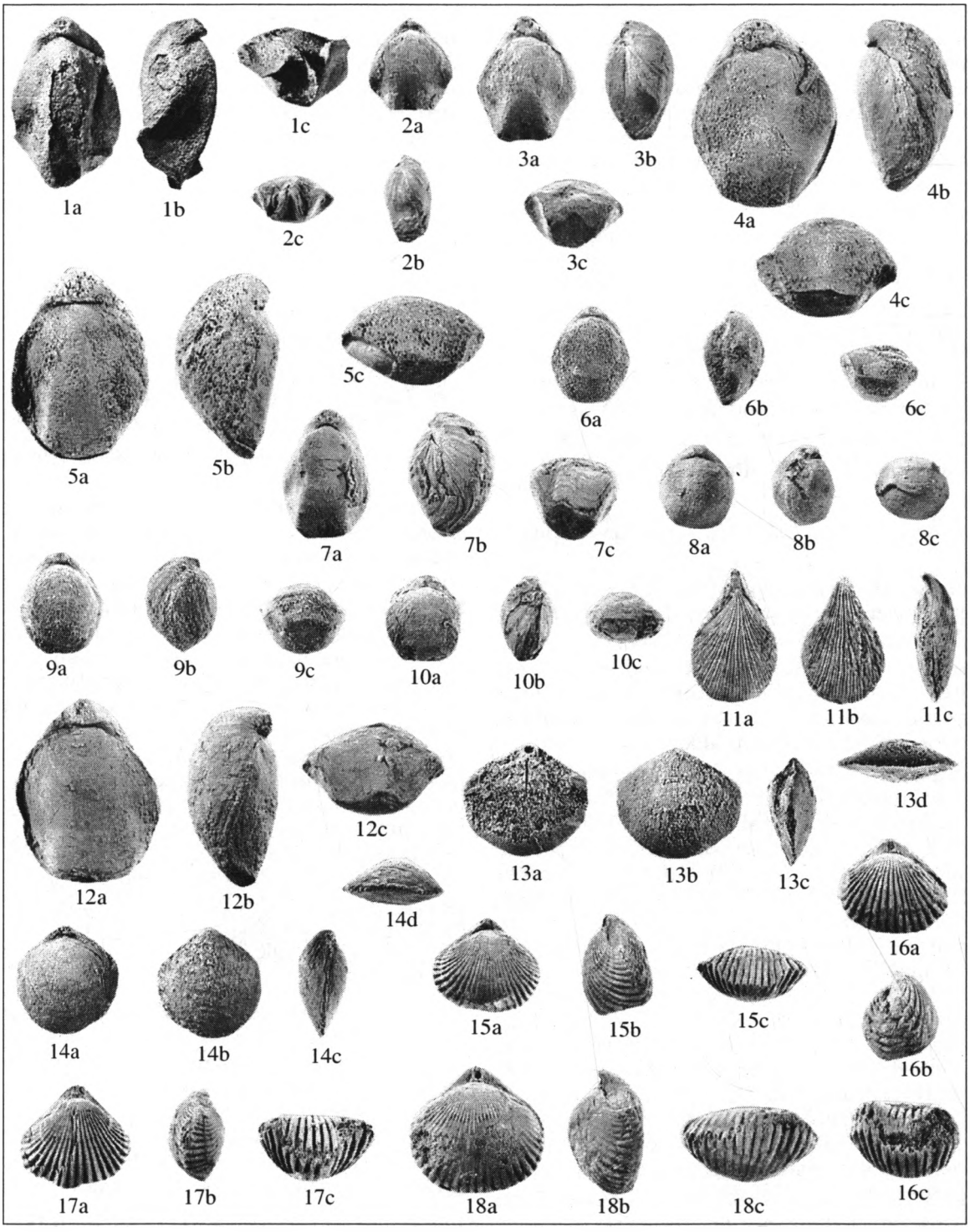

Plate. Valanginian–lower Hauterivian brachiopods of the southwestern Crimea. Specimens 1–9 are from Verkhorech'e section, 10, 11, and 15–18 from Sel'bukhra Mountain, 12 from eastern outskirts of the Nauchnyi Village, 13, 14 from the Patil' Plateau, Prokhladnoe Village area. All figures are in full scale if otherwise not indicated.

(1) *Tropeothyris karakaschi* Smirnova, 1972, 12844/12, Chernyshev Central Research Geologic Prospecting Museum (CCRGP), holotype, dorsal (a), lateral (b), and anterior (c) view, Valanginian; (2) *Loriolithyris carteroniana* (d'Orbigny, 1847), 20/100, Museum of Earth Science (MES), Moscow State University, dorsal (a), lateral (b), and anterior (c) view, lower Valanginian; (3) *Loriolithyris valdensis* (Loriol, 1868), 21/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Valanginian; (4) *Loriolithyris metaitensis* Middlemiss, 1980, 22/100, MES, dorsal (a), lateral (b), and anterior (c) view, upper Valanginian, *Himantoceras trinodosum (Loriolithyris metaitensis)* Zone; (5) *Praelongithyris credneri* (Weerth, 1884), 23/100, MES, dorsal (a), lateral (b), and anterior

missure, lesser thickness, lower beak, and larger apical angle differ them from *Musculina sanctaecrucis* (Gatzigras).

Distribution: lower Hauterivian of the southwestern Crimea.

Superfamily Dallinoidea Beecher, 1893

Family Dallinidae Beecher, 1893

Genus *Terebrataliopsis* Smirnova, 1962

Terebrataliopsis semicircularis Smirnova sp. nov.

Plate, fig. 13.

Name is after Lat. *semicircularis*, semicircular.

Holotype: 3/100, Museum of Earth Science, Moscow State University, southwestern Crimea, Prokhladnoe Village, Patil' Plateau; lower Valanginian.

Material: four well-preserved specimens from the Prokhladnoe Village area, Patil' Plateau.

Description. Wide smooth shell with pteroid lateral slopes near posterior margin is semicircular to rounded-quadrangular in outline. Valves are equally convex. Anterior and lateral commissures are straight. Cardinal margin is long, slightly curved. Ventral valve is most convex in the middle and dorsal one at posterior one-third of the valve length. Greatest width is in the middle. Ventral valve is uniformly convex throughout the length. In some specimens, a poorly pronounced lingula is dorsally directed. Lateral surfaces are flattened. Beak is high, sharpened, slightly curved, and bordered by distinct beak keels. Apical angle is 118°–120°. False area is high and concave. Foramen is large, circular. Pseudodeltidium is low-trapeziform in outline. Dorsal valve convex at posterior half of the shell length has flattened lateral and anterior margins.

Size parameters in mm and their ratios:

Sample	L	W	Th	W/L	Th/L
Holotype 3/100	12.8	14.2	5.0	1.11	0.39
15/100	11.5	12.0	5.8	1.04	0.51
16/100	8.8	10.8	4.5	1.23	0.51
17/100	8.6	9.2	4.1	1.07	0.48

Comparison. In distinction from *Terebrataliopsis quadrata* Smirnova, shells are semicircular, larger in size and wider, having long cardinal margin, equally convex valves, and larger apical angle.

Distribution: lower Valanginian of the southwestern Crimea.

Terebrataliopsis compacta Smirnova, sp. nov.

Plate, fig. 14.

Name is after Lat. *compactus*, compact.

Holotype: 4/100, Museum of Earth Science, Moscow State University, southwestern Crimea, Prokhladnoe Village, Patil' Plateau; lower Valanginian.

Material: three well-preserved specimens from the Prokhladnoe Village area, two from the Sheludivaya Mountain and one from the Patil' Plateau.

Description. Compact, rounded to rounded-pentagonal shells with rounded, slightly projected lateral slopes are slightly greater in length than in width. Ventral valve is somewhat more convex. Anterior commissure is ventrally arched and lateral commissures are straight. Cardinal margin is short, slightly curved. Being widest at posterior one-third of the length, shells are most thick in the middle or slightly closer to beak. Ventral valve is curved roof-like from beak to anterior margin and flattened along margins. Beak is low, slightly curved, bordered by sharply pointed keels. False area is low, concave. Apical angle is 100°–105°. The mesothyridid foramen is large, circular. Pseudodel-

(c) view, upper Valanginian, *Neohoploceras submartini (Loriolithyris metaitensis)* Zone; (6) *Musculina nuciformis* sp. nov., 1/100, MES, holotype, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, ×1.2; (7) *Musculina sanctaecrucis* (Gatzigras, 1948), 24/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone; (8) *Musculina globus* (Pictet, 1872), 12844/17, CCRGP, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian; (9) *Musculina nikischini* sp. nov., 2/100, MES, holotype, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, ×1.5; (10) *Belothyris regularis* Smirnova, 1960, 25/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone; (11) *Symphythyris neocomiensis* (d'Orbigny,1847), 26/100, MES, dorsal view (a), ventral valve (b), lateral view (c), lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone, *Symphythyris neocomiensis* Subzone; (12) *Sellithyris sella lindensis* Middlemiss, 1976, 27/100, MES, dorsal (a), lateral (b), and anterior (c) view, upper Valanginian, *Neohoploceras submartini (Loriolithyris metaitensis)* Zone; (13) *Terebrataliopsis semicircularis* sp. nov., 3/100, MES, holotype, dorsal view (a), ventral valve (b), lateral view (c), anterior view (d), lower Valanginian, ×1.5; (14) *Terebrataliopsis compacta* sp. nov., 4/100, MES, holotype, dorsal view (a), ventral valve (b), lateral view (c), anterior view (d), lower Valanginian, ×1.5; (15) *Cyclothyris rostriformis* (Roemer, 1836), 28/100, MES, dorsal (a), lateral (b), and anterior (c) view, Valanginian–lower Hauterivian; (16) *Cyclothyris walkeri claxbyensis* (Owen et Thurrell, 1968), 29/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone, *Symphythyris neocomiensis* Subzone; (17) *Cyclothyris castellanensis* (Jacob et Fallott, 1913), 30/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone, *Symphythyris neocomiensis* Subzone; (18) *Cyclothyris corallina neocomiensis* (Jacob et Fallott, 1913), 31/100, MES, dorsal (a), lateral (b), and anterior (c) view, lower Hauterivian, *Lyticoceras nodosoplicatum (Musculina sanctaecrucis)* Zone, *Symphythyris neocomiensis* Subzone.

Specimens are collected by Smirnova (1–3, 6, 8, 9, 15), Baraboshkin (4, 5, 7, 10–12, 16–18) and Yanin (13, 14).

tidium is low. Dorsal valve is convex in the posterior half of shell. Wide sinus in the anterior half of dorsal valve is poorly developed.

Size parameters in mm and their ratios:

Sample	L	W	Th	W/L	Th/L
Holotype 4/100	11.5	11.4	6.5	0.98	0.56
18/100	11.0	10.8	5.1	0.99	0.45
19/100	10.2	10.1	4.9	0.98	0.48

Comparison. In distinction from *Terebrataliopsis semicircularis* sp. nov., characteristic of the species are narrower shell, unequally convex valves, ventral curvature of anterior commissure, short cardinal margin, smaller apical angle, and low beak. Rounded shell, position of greatest width, slightly curved cardinal margin, presence of dorsal sinus, and arched anterior margin differ the species from *Terebrataliopsis quadrata.*

Distribution: lower Valanginian of the southwestern Crimea.

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The Eocene Fossil Plants, Spores, Pollen, and Dinocysts from the Tavda Formation, the Pavlodar Area near the Irtysh River

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Abstract—The Bartonian Age of subtropical small-leaved fossil flora from green clay members of the Tavda Formation, the Pavlodar area near Irtysh River, is substantiated for the first time based on dinocysts detected in outcrops of the Irtysh–Karaganda Channel. The microplankton composition suggests that the lower Tavda Subformation was deposited during the first transgressive phase of the late Eocene. A sea basin that spread at that time over southern and central regions of West Siberia was connected with the Tethys and reached the northern marginal areas of the Kazakh Hills. The *Castanopsis* pollen is dominant in palynological assemblage and occurs in association with fine pollen of the *Quercus gracilis–Q. graciliformis* group. The assemblage is likely indicative of the earliest phase of climatic inversion, when the Lutetian monsoon climate in middle latitudes of western Asia turned into seasonal climate of the Mediterranean type with hot summers and most intense precipitations during the winter period.

Key words: flora, dinocysts, spores, pollen, Paleogene, Pavlodar area of the Irtysh River, West Siberia, palynological zone.

INTRODUCTION

Geological Background

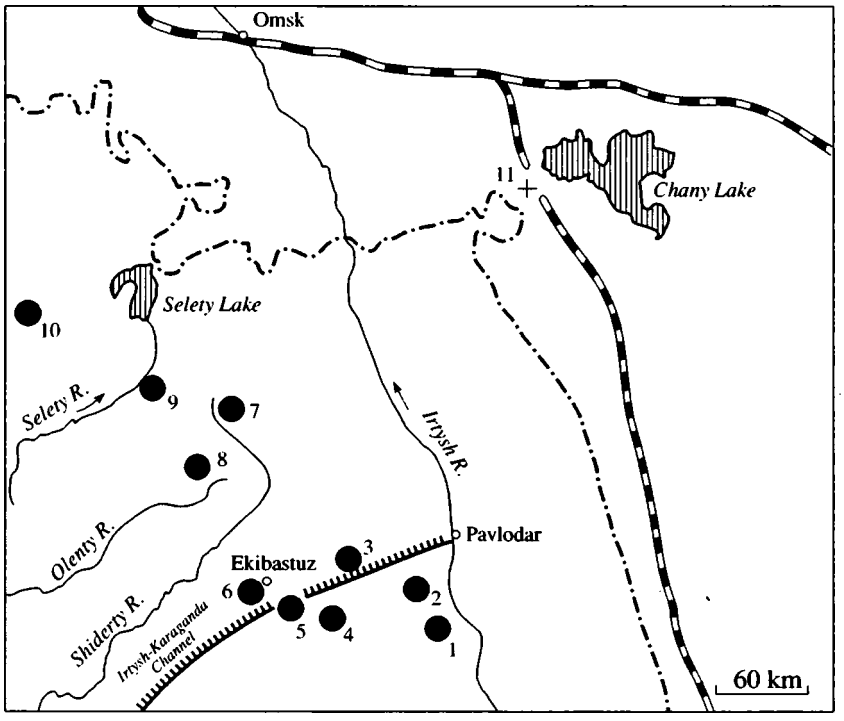
The Paleogene flora localities are known from several geological structures of the study region: (1) near the boundary between the circum-Caspian Syneclise and Turan plate, from the Emba River to southern spurs of the Urals; (2) in the northern to northeastern flank of the Kazakh massif facing the West Siberian plate (Pavlodar area of the Irtysh River); (3) in the northern Aral area, Ustyurt, and Turgai depression; and (4) in the Zaisan basin (Makulbekov, 1997). In three former cases, their ages are inferred from general evolutionary trends of Paleogene floras in middle latitudes of the Northern Hemisphere and from correlation of continental flora-bearing deposits with marine sediments, the age of which is established based on fossils of a higher stratigraphic resolution, primarily on microplankton. In the Zaisan basin isolated from seas, and partially in the Turgai depression, dating of floras is corrected with due account for concurrent mammal remains, herpetofauna, mollusks, and other representatives of terrestrial biota.

This work is dedicated to floras from a section of the Karasor Lake depression in the Pavlodar area (Fig. 1). Antypko (1959) who carried out the geological survey in that area and afterward Zal'tsman (1968) arrived at the conclusion that paleogeography of the area was under control of recurrent transgressions and regressions, like in the southern margin of the West Siberian plate. During the transgressions, seawaters flooded the Kazakh Hills, while sandy beds with plant remains accumulated in exsiccation areas during regressions. Accordingly, it is necessary to analyze in detail the transgressive–regressive rhythms in the basin and their periods in order to date reliably the flora-bearing deposits.

During the late Lyulin Vor and Tavda times, the lake area was recurrently flooded by sea, being a part of the sea coastal zone during regressions. Melkozem and abundant plant remains, leaves and sprouts, have been transported to that coastal zone. Floral remains are buried in both the seaward clay facies (sprouts, stems, and frutifications of Taxodiaceae) and sandy deposits of shoals and beech zone (Makulbekov, 1972, 1997). In distinction from the Zhamantuz, Takyrsor, and other localities in the Pavlodar area, where flora-bearing beds rest directly on pre-Cenozoic weathering crust or on older magmatic rocks, these beds are intercalated either with the lower–middle Eocene clays and siliciliths of the Lyulin Vor Formation, or with the middle–upper Eocene clays of the Tavda Formation in the Karasor depression sections.

Flora-Bearing Section

The studied locality of large plant remains and microfossils is exposed in walls of the Irtysh–Karaganda Channel (IKC) northwestward of the Kalka-



Main localities of the Eocene floras in the Pavlodar area, the Irtysh River basin: (1) Zhamantuz; (2) Takyrsor; (3) Kudaikol'; (4) Karasor; (5) IKC; (6) Ekibastuz; (7) Shiderty; (8) Olenty; (9) Selety; (10) Kulboldy; (11) Borehole 9, Chistoozernoe Settlement (Chany Lake).

man Station near the Karasor Lake that represents a spacious depression down to 80 m deep. The lake is incised in marine deposits underlying the younger flora-bearing quartzose sandstones, which armor the plain surface above the lake. Walls of the channel excavated along the southern edge of the lake depression exhibit strata of marine deposits, which are a few tens of meters thick and underlie the flora-bearing beds, locally alternating with them. The diverse evergreen flora of the lower Lutetian has been described from lower beds of this area exactly (Makulbekov, 1972). The entire leaves of *Castanopsis kazachstanensis* Makul., *C. pavlodarensis* Makul., and of genera *Lithocarpus, Dryophyllum, Viburnum, Sterculia*, and *Palmophyllum* are dominant in this flora that differs from older late Thanetian and Ypresian floras of Kazakhstan, those of the nearby Ekibastuz and Maisk localities (Pavlodar area) included, because it is lacking representatives of the genus *Ushia*. Within the marine facies of the upper Lyulin-Vor Subformation, the flora-bearing Karasor Beds correspond to a sand member that separates the green clay member with fish remains (the Nyurol'ka Formation in nomenclature of N.B. Shatskiĭ) from the underlying clayey–siliceous member, which yields a diverse assemblage of the early Eocene microfossils. These relations are consistent with paleontological data on borehole section near the Ekibastuz–Pavlodar railway crossing the Karasor depression (Kopytova *et al.*, 1960; Zal'tsman, 1968).

In the same depression, though a bit westward of the main Karasor locality (closer to the Shiderty River), the channel walls exhibit the higher strata of marine genesis, which correspond to the green clay member of the Tavda Formation, which yields fossil plant remains. The member is overlain in turn by light-colored brownish gray sandstones, siltstones, and clays of the Tavda type. The IKC flora from these rocks includes both the evergreen and deciduous taxa. The flora-bearing section exposed in the striped wall 15 km northwestward of the Kalkaman railway station consists of the following units (from the base upward):

(1) Gravel and pebble beds (apparent thickness over 2 m);

(2) Dark green thick-bedded compact clay with thin interlayers of gray sand; the rocks enclose fragments of lignified trunks of *Taxodioxylon sequoianum* (Mercklin) Schmal. preliminarily identified by I.A. Shilkina (3.2 m thick);

(3) Dark green (brown when dried) bedded clay with horizontally-lying coalified trunks of *Taxodioxylon* sp. (3.1 m thick);

(4) Dark green (pinkish gray when dried) fine-laminated clay; near the base there are buried sprouts of Taxodiaceae, wood fragments, and impressions of stems and leaves of dicotyledons and Gramineae. Palynological sample 1 is from this unit (3.2 m thick).

Beds occurring above, maybe after a hiatus, are as follows:

(5) Yellow sand with clay pellets derived from underlying rocks (0.2 m thick);

(6) Green grumous clay and gray silt with plant remains; palynological sample 2 is collected here (1.5 m thick);

(7) Farinaceous siltstone with lenticular intercalations or rusty and orange sands (2.5 m thick);

(8) Soil formed after weathering debris of gray to yellowish and orange flaggy quartzose sandstones with leaf impressions (0.5 m thick).

Well-preserved leaves and sprouts of arboreal plants have been collected from the upper part of the sections (units 6–8).

RESULTS AND DISCUSSION

Fossil Plants

The fossil flora from dark green clays of the Tavda Formation (Bed 4) consists predominantly of *Sequoia couttsia* and *Taxodium* sprouts mixed with remains of small evergreen taxa *Litsea primigenia* (Ung.) Takt., *Leucothoe protogaea* (Ung.) Schimp., *Quercus pseudoneriifolia* Vikul., *Myrica lignitum* (Ung.) Sap., *M. acuminata* Ung., and *Dryophyllum curticellense* (Watel.) Sap. et Marion. The associated rare deciduous forms are *Sassafras ferritianum* Massal., *Liquidambar europaea* A. Br., *Zelkova zelkovifolia* (Ung.) Buzek et Kotlaba, and *Comptonia acutiloba* (Heer).

In the upper section part (beds 6–8), flora is mostly large-leaved mesophilic, dominated by deciduous taxa *Comptonia acutiloba, Alnus* sp., and *Populus* sp. In addition to entire *Quercus* leaves, it includes dentate leaves of this genus as well. *Myrica* species, *Liquidambar*, and *Taxodium* transit at this level from the older assemblage.

Palynological Characterization

Two diverse assemblages of palynomorphs, one from the green clay with plant remains (Bed 4, Sample 1) and the other one from similar flora-bearing clay overlying sands (Bed 6, Sample 2), have been analyzed by Zaporozhets. In addition to spores and pollen, the lower sample yielded the diverse organic-walled phytoplankton (dinocysts, acritarchs, and prasinophytes) specifying age of the older plant microfossils. The upper sample is almost lacking the organic-walled phytoplankton, but the diverse spores and pollen from the sample clarify the flora composition and age.

Abundance of palynomorphs in both samples is sufficient for statistic analysis (921 and 396 specimens in samples 1 and 2 respectively). In Sample 1, spores and pollen represent 80.3% of all palynomorphs (34.8% of angiosperms and 45.5% of gymnosperms). The organic-walled phytoplankton (19.7% in total) consists of dinocysts (13.2%) and of acritarchs and prasinophytes (6.5% in sum). Percentages of different palynomorph groups from Sample 2 are as follows: 39.1% of angiosperms, 54.5% of gymnosperms, and 7.4% of prasinophytes. Two damaged specimens of dinocysts found in this sample (*Cordosphaeridium* and *Chytroeisphaeridia*) are likely redeposited.

Dinocysts from Sample 1 are diverse and indicative of the open sea environment and normal water salinity (Plates I and II). Dominant in this assemblage of 55 taxa are *Phthanoperidinium* (*Phth. spackmansii, Phth. geminatum, Phth.* cf. *eocenicum, Phthanoperidinium* sp.), *Wetzeliella* (*W. irtyschensis, W. articulata, Wetzeliella* sp.), *Charlesdowniea–Kisselovia* group (*Ch. tenuivirgula, K. ornata*), *Thalassiphora* ex gr. *turgaica-elongata, Heteraulacacysta* (*H. pustulata, H.* cf. *campanula, Heteraulacacysta* sp. 1 that is likely a new form), and *Glaphyrocysta* (*G. semitecta, G. exuberans*). Taxa listed above coexist with single specimens of *Areosphaeridium diktyoplokum, Achomosphaera* sp., *Spiniferites* sp., *Samlandia* sp., *Lingulodinium* sp., *Hystrichokolpoma rigaudiae, Hystrichokolpoma* sp., *Deflandrea phosphoritica, Deflandrea* sp., *Cerodinium markovii, Cerodinium* sp., *Alterbidinium* sp., *Rhombodinium* sp., (?)*Batiacasphaera* sp., (?)*Spinidinium* sp., *Achilleodinium biformoides, Achilleodinium* sp., (?)*Chytroeisphaeridium* sp., *Pteridium* sp., *Hystrichosphaeropsis* sp., *Kallosphaeridium* sp., *Homotryblium oceanicum, H. pallidum, Homotryblium* sp., *Oligosphaeridium* sp., *Cordosphaeridium exilimurum, Cordosphaeridium* sp., *Horologinella incurvata, Horologinella* sp., *Phelodinium* sp., *Adnatosphaeridium robustum*, and *Adnatosphaeridium* sp. Among prasinophytes and acritarchs, there are 7 taxa at least. Main of them are *Paucilobimorpha* forms (*P. triradiata, P. granuligera, Paucilobimorpha* sp.), while *Micrhystridium* and *Crassosphaera* species are less abundant. Single specimens represent *Ptanospermella* and *Tasmanites* cf. *Lanjaria.*

Spores and pollen from Sample 1 are similarly diverse (Plate II). The palynological spectrum mostly consists of Taxodiaceae (21%), *Pinus* (9%), and Fagaceae (24%). Pollen of *Quercus graciliformis* represents over 50% among Fagaceae. The abundance rate of 1 to 4% is characteristic of *Glyptostrobus* sp., *Taxodium* sp., *Quercus gracilis, Q. conferta, Q. sibirica, Quercus* sp., *Castanopsis pseudocingulum, Castanopsis* sp. (all *Castanopsis* forms up to 4% in sum), *Alnus* sp., *Pterocarya* sp., *Rhus* sp. (*R. coriariformis* included), *Nyssa* sp., *Comptonia* sp., Hamamelidaceae–Altingiaceae (up to 10% in sum, Hamamelidaceae gen. indet., *Hamamelis scotica, Hamamelis* sp., *Liquidambar* sp., *Fothergilla gracilis, F. vera, Fothergilla* sp., *Altingia* sp., *Corylopsis princeps*, and *Corylopsis* sp. included), *Ilex* sp., *Aralia* sp., Araliaceae gen. indet., and *Tricolporopollenites* sp. (*C. edmundii* and *C. liblarensis* included). Single pollen grains are those of *Picea* sp., *Tsuga* sp., *Keteleeria* sp., *Sciadopitys* sp., *Ephedra cheganica, E. fusiformis, Ephedra* sp.,

Plate I

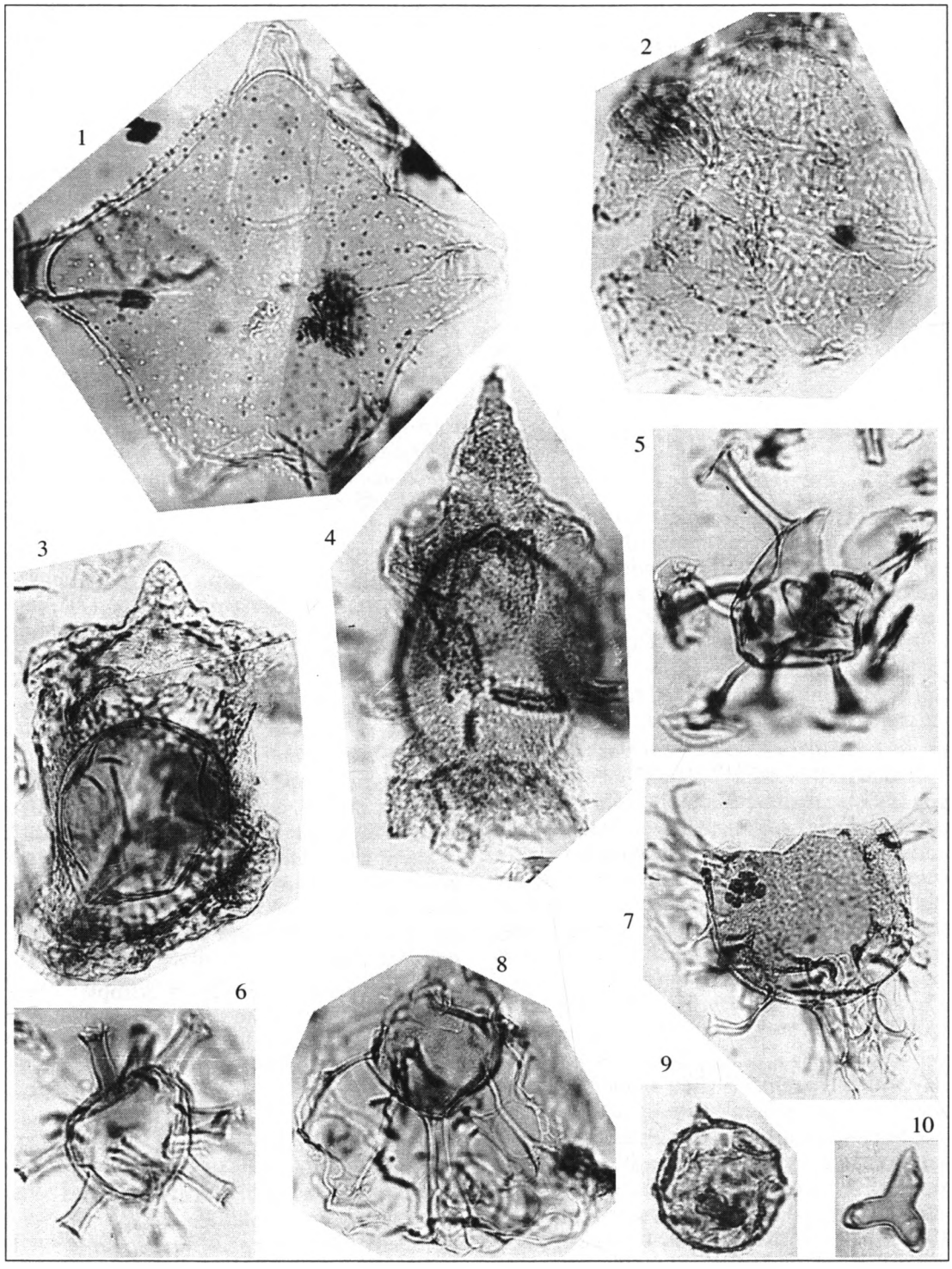


Plate I. Organic-walled phytoplankton from Sample 1 (the middle Eocene of the Pavlodar area near the Irtysh River; magnification 500 in figs. 3, 5, 7, 8 and 750 in the others):

(1) *Wetzeliella irtyschensis* Alberti; (2) *Kisselovia ornata* Vozzhennikova; (3) *Thalassiphora elongata* Vassilieva; (4) *Araneosphaera araneosa* Eaton; (5) *Areosphaeridium diktyoplokum* (Klumpp) Eaton; (6) *Homotryblium tenuispinosum* Davey and Williams; (7) *Glaphyrocysta ordinata* (Williams and Downie) Stover and Evitt; (8) *Adnatosphaeridium robustum* (Morgenroth) de Coninck; (9) *Phthanoperidinium eocenicum* (Cookson and Eisenack) Lentin and Williams; (10) *Paucilobimorpha triradiata* de Coninck.

Plate II

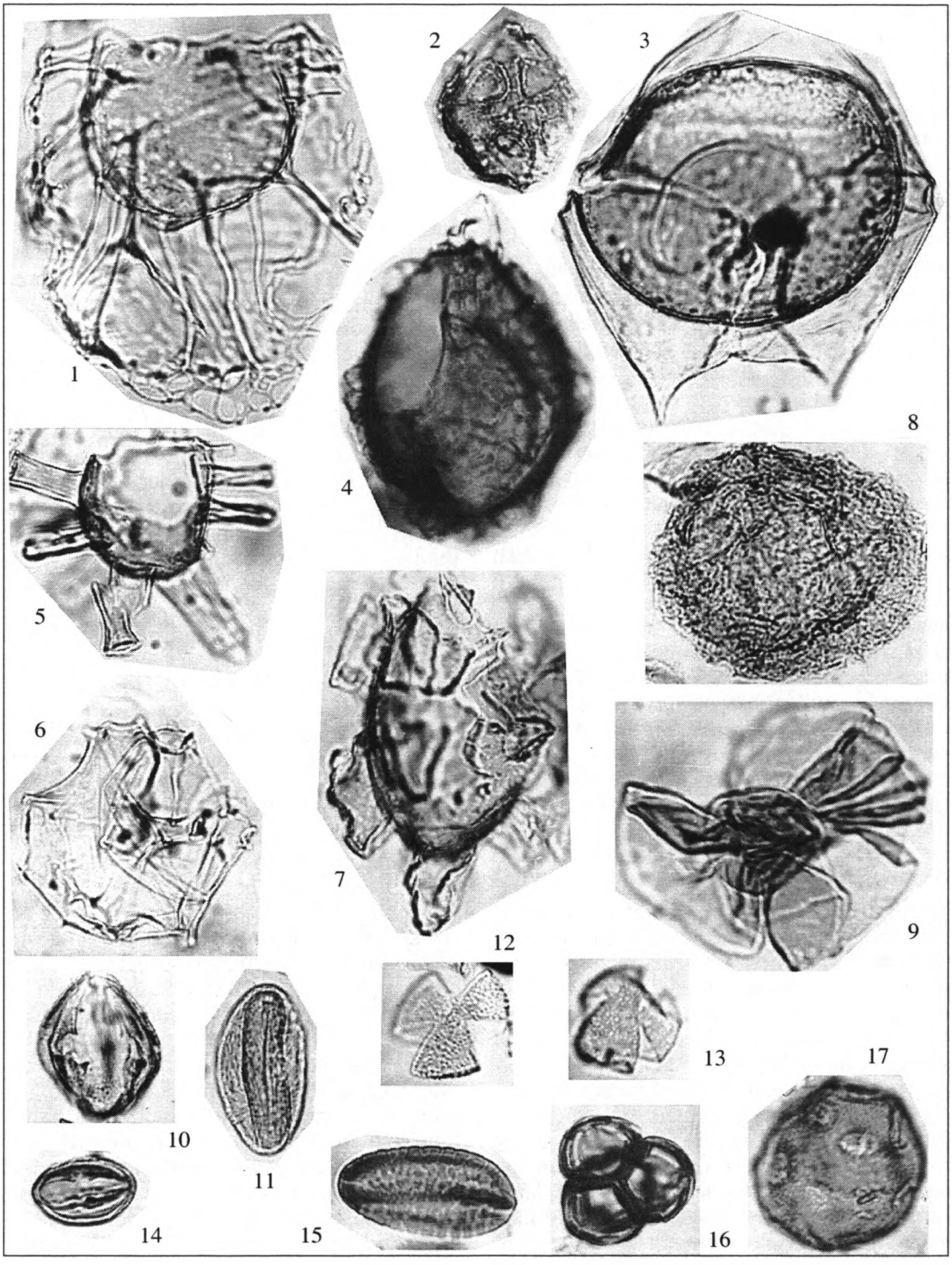


Plate II. Organic-walled phytoplankton and pollen from Sample 1 (the middle Eocene of the Pavlodar area near the Irtysh River; magnification 1000 in figs. 12–14 and 750 in the others):

(1) *Glaphyrocysta semitecta* (Bujak) Lentin and Williams; (2) *Phthanoperidinium stockmansii* (de Coninck) Lentin and Williams; (3) *Deflandrea phosphoritica* Eisenack; (4) *Samlandia chlamidophora* Eisenack; (5) *Hystrichokolpoma salacia* Eaton; (6) *Hystrichosphaeropsis* sp.; (7) *Achilleodinium biformoides* (Eisenack) Eaton; (8) *Heteraulacacysta* sp. 1; (9) *Ptanospermella* sp.; (10) *Nyssa* sp.; (11) *Quercus graciliformis* Boitzova; (12) *Fothergilla gracilis* Lubm.; (13) *Hamamelis scotica* Simpson; (14) *Rhus ustjurtensis* Boitzova; (15) *Ephedra cheganica* Shakhm.; (16) Ericaceae gen. indet; (17) *Liquidambar* sp.

Betula sp., *Carpinus* sp., *Corylus* sp., *Carya* sp., *Engelhardtia* sp., *Juglans sieboldianaeformis, Juglans* sp., *Platycarya* sp., *Castanea* sp., *Fagus* sp., *Tilia* cf. *sulcata, Tilia* sp., *Crewia* sp., *Myrica* sp., Myricaceae gen. indet., *Ulmus* sp., Ulmaceae gen. indet., *Platanus* sp., Sapotaceae gen. indet., Ericaceae gen. indet., *Magnolia* sp., *Sparganium* sp., *Salix* sp., Rosaceae gen. indet., Oleaceae gen. indet., *Elaeagnus* sp., *Sterculia* sp., Moraceae gen. indet., *Sabal turgaica*, and Palmae gen. indet. Single representatives of *Sphagnum* and *Osmunda* sp. are established among spores.

Palynological spectrum of the upper flora-bearing bed is essentially different. As is already noted, phytoplankton of the bed is represented by green algae (7.4%), mainly by *Crassosphaera* and single specimens of *Ptanospermella, Tytthodiscus*, and *Cymatiosphaera.* In addition to Pinaceae representing about 38%, dominant in palynological spectrum are flowering plants, such as *Quercus* (mostly evergreen species *Quercus gracilis, Q. graciliformis, Q. conferta*, and *Q. sparsa*), *Liquidambar* sp., *Myrica* sp., and *Nyssa* sp., (from 7 to 9%). *Pterocarya* sp., *Ulmus* sp., *Comptonia* sp., *Ilex* sp., *Tilia* sp., and Ericaceae gen. indet. constitute 3 to 6% of the spectrum. Single (1 to 5) pollen grains characterize the abundance rate of *Picea* sp., Taxodiaceae gen. indet., *Castanea* sp., *Fagus* sp., *Platycarya* sp., *Juglans* sp., *Engelhardtia* sp., *Alnus* sp., Hamamelidaceae gen. indet., *Corylopsis* sp., *Hamamelis* sp., *Fothergilla* sp., *Salix* sp., *Celtis* sp., Ulmaceae gen. indet., *Rhus* sp., Myricaceae gen. indet., *Grewia* sp., *Palmae* sp. (= *Trachycarpus* sp.), Oleaceae gen. indet., Chenopodiaceae gen. indet., *Parthenocissus* sp., Rosaceae gen. indet., and *Triatriopollenites* sp.

DISCUSSION

As for stratigraphic position of Sample 1 and corresponding clay bed, it seems to be close, if not identical, to that of the green clay member penetrated by Borehole 9 near the Chistoozernoe Settlement northward of the Karasor Lake (the southwestern termination of the Chany Lake, Fig. 1). Multidisciplinary paleontological study of that borehole section specified ages of particular stratigraphic intervals and revealed breaks in sedimentation at the levels of sandy members. According to regional drilling data (Zal'tsman, 1968), the sandy member considerably grows in thickness toward the depression periphery and show the successive replacement of marine facies first by coastal-marine and then by continental flora-bearing sediments. The studied assemblages on microplankton, especially of planktonic foraminifers and dinocysts (Beniamovskiĭ *et al.*, 2002), imply that breaks in sedimentation at basal levels of sandy members of the upper Lyulin Vor Subformation, which were penetrated by Borehole 9, correspond to the early Ypresian, terminal Ypresian–initial Lutetian, initial late Lutetian, and early Bartonian (the Tavda Formation base). During the more intense regression of the Priabonian (Tavda) time, marine sediments became replaced in the easterly direction by coastal-marine and then by continental facies of the Yurkovo Formation.

Species *Areosphaeridium diktyoplokum, Kisselovia ornata, Wetzeliella irtyschensis, W. articulata, Charlesdowniea tenuivirgula*, and *Thalassiphora elongata* present among dinocysts and *Paucilobimorpha* forms occurring among acritarchs represent distinctive features of organic-walled phytoplankton from clays of the lower Tavda Subformation penetrated by Borehole 9 in the depth interval of 420–390 m. Also remarkable is a great diversity of *Phthanoperidinium* forms characteristic of the Tavda Formation in general and of its lower subformation in particular. In the spores and pollen assemblage from the same depth interval of Borehole 9, taxa of flowering plants dominate over gymnosperms. Pollen of *Pinus* and Taxodiaceae is of approximately equal abundance among the latter, while *Quercus, Castanopsis*, and Hamamelidaceae are most important representatives of the former. The aforementioned assemblage of phytoplankton is very close in composition to Bartonian assemblages known in West Siberia, the East European platform, Turan plate, and Crimea–Caucasus region. It is analogous as well to concurrent assemblages of Central and Western Europe (*Areosphaeridium diktyoplokum–Rhombodinium draco* and *Rhombodinium porosum* zones). Unfortunately, the Bartonian index species *A. diktyoplokum* and *Rhombodinium porosum* occur as single specimens only in the studied Sample 1 from flora-bearing beds of the Pavlodar area near the Irtysh River. As for the Priabonian species *Charlesdowniea clathrata angulosa*, it has not been encountered at all. Consequently, the corresponding beds can be confidently attributed to the Bartonian, maybe to its upper interval, as we can judge from presence of the late Bartonian *Rhombodinium porosum.* This is consistent with presence of *Acarinina* sp. and *Catapsydrax* forms at the same level of the lower Tavda Subformation penetrated by Borehole 9 (Beniamovskiĭ *et al.*, 2002). Sample 2 and Bed 6 that is lacking dinocysts may be correlated with the higher stratigraphic interval of Borehole 9 (390–330 m below surface), where clay is substituted by silt and sand. This interval is of the Priabonian age according to the composition of studied palynomorphs. Characteristic of dinocysts from this interval are the mass abundance of *Hydropteris* and *Azolla* forms and appearance of the subspecies *Charlesdowniea clathrata angulosa.* Like in Sample 1, pollen of *Pinus* is still rather abundant here, and representative pollen of thermophilic plants partially belongs to evergreen taxa: oaks of the *Quercus gracilis–Q. graciliformis* group, Hamamelidaceae–Altingiaceae (*Fothergilla, Corylopsis, Hamamelis*, and *Liquidambar*), *Nyssa, Rhus, Grewia, Palmae (Trachycarpus)*, and Ericaceae. The Priabonian age of the upper flora-bearing bed is also consistent with the mixed composition of fossil plant remains. Identified among the latter are evergreen, typical subtropical and deciduous mesophilic taxa *Myrica lignitum* (Ung.)

Sap., *Liquidambar europaea* A. Br., *Zelkova zelkovifolia* (Ung.) Buzek et Kotlaba, *Dryophyllum curticellense* (Watel.) Sap. et Marion, Lauraceae, and others.

CONCLUSION

Correlation

Described below is the most probable scheme of correlation between the flora-bearing Eocene beds of the Pavlodar area near the Irtysh River and rudaceous marine deposits in the south of the West Siberian plate. The scheme specifies ages of the studied Paleogene floras, the stratigraphic succession of which was originally outlined by Makulbekov (1997).

(1) The coarse- to medium-grained sandstone members of the Ekibastuz and Maisk sections, which yield rare *Sequioa* in association with *Litsea, Persea, Laurus* (Lauraceae), and small xeromorphic *Ushia*, appear to be correlative with the middle Ypresian member of glauconite sandstones at the base of the upper Lyulin-Vor Subformation. In Borehole 9, similar sandstones likely underlie the Ypresian clay that is attributed to the *Dracodinium varielongitudum* Zone of the dinocyst scale (Beniamovskiĭ *et al.*, 2002). This basal sandstone member of the upper Lyulin-Vor Subformation is distinguishable not only in the Barabinsk lithofacies zone northward of Pavlodar, but also in the easterly Kulunda zone (Zal'tsman, 1968).

(2) The flora-bearing sandstones of the Karasor–Kudaikol' area, where they yield remains of evergreen large-leaved Fagaceae with entire leaves (*Castanopsis kazachstanensis* Makul. and other species of the genus, *Litliocarpus, Quercus, Dryophyllum, Viburnum, Sterculia*, and *Palmophyllum*), grade in lateral direction into the glauconite sandstones underlying the green clay member with ichthioliths ("Fish Clay"). This clay member corresponds to the upper part of the upper Lyulin-Vor Subformation above the gray clay–siliceous member. In Borehole 9, the lower beds of green clay bear dinocysts of the *Wetzeliella articulata–Systematophora placacantha* Zone, the *Castanea crenataeformis–Castanopsis pseudocingulum* palynological assemblage, and planktonic foraminifers of the middle Lutetian *Acarinina rotundimarginata* Zone (Beniamovskiĭ *et al.*, 2002).

(3) Cavernous sandstones, which enclose the xeromorphic small entire leaves of *Quercus takyrsorianus* Makul., Lauraceae, Ericaceae (*Laurus primigenia, Leucothoe protogaea, Lomatia firma*), and conifer remains (*Sequoia, Pinus*) in the Takyrsor, Olenty, Selety, Kulboldy, and Zhamantuz localities, are widespread in many areas of Kazakhstan beyond the Irtysh River basin. Characteristic of them is palynological assemblage predominantly consisting of angiosperm taxa, *Castanopsis* and *Quercus gracilis–Q. graciliformis* group included. These sandstones were deposited during a remarkable shoaling stage of the West Siberian sea evolution, when similar and even coarser glauconite sandstones with gravely and pebbly inclusions indicative of the underlying strata erosion (the log horizon "T" at the Tavda Formation base) accumulated in a vast area. In Borehole 9, the lower Bartonian sandstones under consideration, which are discriminated into the *Paucilobimorpha–Micrhystridium* Beds based on characteristic taxa of organic-walled phytoplankton, overlie the uppermost Lyulin-Vor strata bearing planktonic foraminifers of the *Hantkenina alabamensis* Zone (transitional beds from the terminal Lutetian to Bartonian). All the fossil plant remains have been buried in a seaside plain, since remains of *Macclintockia* that is a plant populating coastal areas of sea basins are present in the Zhamantuz flora (Budantsev, 1956; Makulbekov, 1972).

(4) Green clays with remains of the Bartonian small-leaved subtropical flora that is described in this work have also counterparts among facies deposited in proximity to the open-sea settings. They are directly correlative with clays of the lower Tavda Subformation penetrated by Borehole 9, which yield palynomorphs of the same *Quercus graciliformis–Castanopsis crenataefolium* palynological assemblage and dinocysts of the *Rhombodinium porosum* Zone.

(5) Flaggy sandstones with the Akzhar and Pavlodar mesophilic floras of the mixed type (Taxodiaceae, entire-leaved evergreen Lauraceae, *Leucothoe, Quercus* ex gr. *Q. pseudoneriifolia, Dryophyllum*, and diverse deciduous taxa, such as *Liquidambar europaea, Zelkova zelkovifolia, Comptonia acutiloba, Alnus, Ulmus*, and others) correspond to the regressive sandstone–siltstone member in the middle part of the Tavda Formation (the log horizon "B", *Charlesdowniea clathrata angulosa* of the dinocyst scale, and palynological *Quercus gracilis–Quercus graciliformis* Zone of the Priabonian). The same floral remains characterize continental deposits of the Yurkovo Formation that replaces the Tavda Formation in the eastern periphery of the West Siberian plate.

Climate

The succession of floras considered above reflects climatic changes in northern Kazakhstan and southern West Siberia during the Eocene epoch.

During the early Eocene and in the second half of the Lutetian, the seasonally humid climate of the monsoon type was characteristic of both aforementioned regions. This is evident from predominance of the genus *Castanopsis*, over 100 species of which are at present the characteristic representatives of subtropical floras, in the Karasor assemblage of plant megafossils and in palynological spectra. The absolute majority of recent *Castanopsis* forms populates now the regions of the monsoon climate (southern and southeastern Asia). Some other basic and characteristic components of the Karasor flora, for instance, Lauraceae (*Persea, Cinna-*

momum) and associated Sterculiaceae and Magnoliaceae, also suggest the monsoon character of climate.

In the Lutetian–Bartonian boundary period, when the epicontinental West Siberian sea was losing connections with the Arctic basin, the monsoon climate gave place to the other seasonal climate close to that of the present-day Mediterranean region, where the bulk of precipitation is confined to the winter period. This climate was characteristic of both the Bartonian and Priabonian ages, when small-leaved xeromorphic oaks *Quercus takyrsorianus* Makul. and species of the section Protobalanus Trel. (*Quercus chrysolepis* Liebm., *Q. virginiana* Mill, and others) became dominant in the vegetation cover. Recent analogues of these taxa are widespread now in Mexico and along the Californian coast of the United States (Makulbekov, 1972), where summer seasons are arid. Oaks of this kind (*Quercus ilex* L., *Q. cocciferae* L.) are typical as well of the Mediterranean region. All the listed oak taxa populate at present the dry slopes and are characteristic components of the Californian chapporals and Mediterranean maquis. *Ephedra* is a common taxon of the Bartonian palynological spectra. By the end of the Eocene, when cooling was in progress, the small-leaved xerophilous flora gave place to the large-leaved mesophilic one, and climate of the Pavlodar area near Irtysh River lost the Mediterranean features and turned into the moderately warm one, displaying a more regular annual distribution of precipitates.

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Onset of Connections between the Pacific and Arctic Oceans through the Bering Strait in the Neogene

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Abstract—New data on the Neogene diatoms, mollusks, and magnetostratigraphy in the North Pacific and surrounding structures suggest the Bering Strait incipient opening in the latest Miocene or probably in the earliest Pliocene. Stratigraphic positions of oldest Arctic migrants in the Pacific sequences and of Pacific migrants in the Arctic sections are defined. Tectonic processes and eustatic fluctuations as possible causes of the strait opening are analyzed. The consequent changes in biota migration directions, oceanic circulation, and bioproductivity of planktonic groups are discussed.

Key words: Neogene, Bering Strait, diatoms, mollusks, oceanic connections, biota migration.

INTRODUCTION

The Neogene history of the Bering Strait has been considered in many publications. Pioneering works appeared in the 1960s–1970s, when new data on the Cenozoic of Iceland and the North Atlantic were obtained (Durham and MacNeil, 1967; Einarsson *et al.*, 1967; Strauch, 1970, 1972; Einarsson and Yu. Gladenkov, 1973; MacNeil, 1973; Yu. Gladenkov, 1974, 1976, 1978). In the next period of the late 1970s–1990s, the Neogene stratigraphy of Kamchatka, Japan, and Alaska was studied in detail (Allison, 1978; Yu. Gladenkov *et al.*, 1991a; *Detailed...*, 1992; Suzuki and Akamatsu, 1994; Ogasawara, 1998). The third information wave occurred at the turn of the century after Alaska sections had been restudied (Marincovich and A. Gladenkov, 1999, 2001; Marincovich, 2000; Marincovich *et al.*, 2002; A. Gladenkov *et al.*, 2002, A. Gladenkov, 2003). It is natural that views on the Neogene history of the strait and on its opening time changed during the last 35–40 years, but both problems remain challenging nowadays.

Yu. Gladenkov (1995) outlined seven solved and six unsolved problems of the paleo-Bering Strait history. Now, nearly ten years later, we can discuss them again and correct the strait development model. The latest data on Neogene diatoms (first of all), mollusks, and magnetostratigraphy of Kamchatka, Alaska, and Japan enable a more reliable dating and correlation of many formations of the North Pacific region. In turn, new stratigraphic interpretations elucidate the appearance time of the characteristic molluscan assemblage with *Astarte* in the Pacific sections. In opinion of many geologists, the assemblage penetrated into the North Pacific from the Arctic basin at the time of the Bering Strait opening. The latter event has been also considered long as responsible for the appearance of Pacific molluscan assemblage with *Serripes groenlandicus* in Neogene deposits of the North Atlantic.

In addition to new regional materials, geochronology of general standard zonations has been more precisely correlated with the geomagnetic polarity time scale (Berggren *et al.*, 1995). The modern definition of the Neogene stage and zone boundaries eliminated previous uncertainties in stratigraphy of corresponding sedimentary sequences in the Pacific areas. The revised time spans of particular stratigraphic levels corrected essentially the previous correlation schemes (for instance, the age correction of some levels in the well-known Karaginskii section of northeastern Kamchatka is as great as 0.5 m.y.). As a result, the newly obtained geological data enable a more precise correlation of tectonic, climatic, and biotic Neogene events in the northwestern and northeastern Pacific.

Present-Day Bering Strait as a Seaway

At present, the water exchange between the Pacific and Arctic oceans takes place through the Bering Strait (Fig. 1) owing to the main northward current and a minor countercurrent in its western part. The northward inflow of the North Pacific water into the Arctic basin is as great as 10^6–15^6 m^3/s (Shaffer and Bendtsen, 1994; Reason and Power, 1994). The Bering Strait is 96 km long, at least 86 km wide, and about 50 m deep. In spite of these comparatively small dimensions, the strait transports every year about 50 million tons of suspended terrigenous matter and 60 million tons of suspended biogenic matter from the Pacific basin into the Chukchi Sea (Pavlidis, 1982). The biogenic matter,

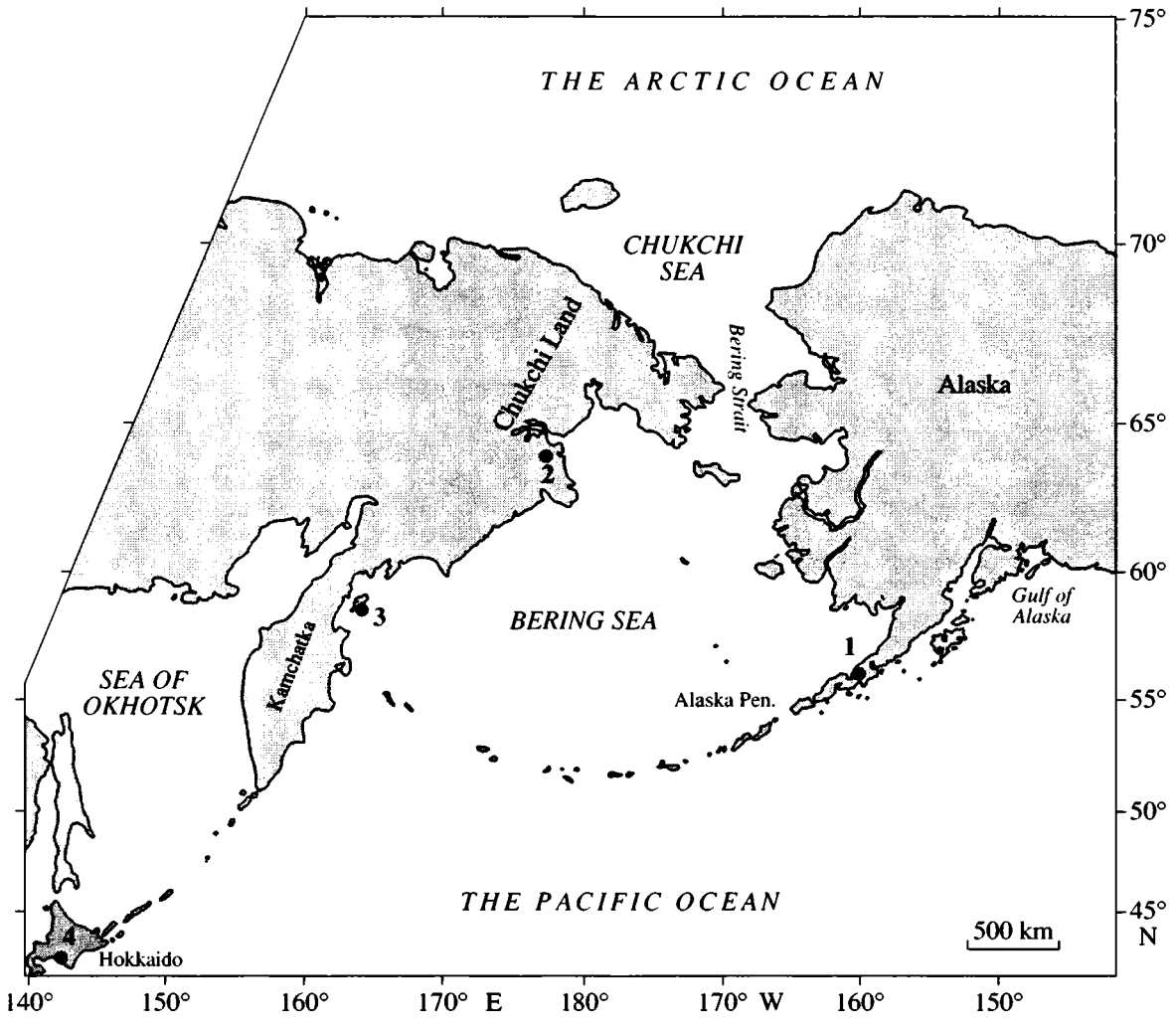

Fig. 1. The Neogene sections with the oldest *Astarte* (black circles) in the North Pacific region: (1) Alaska Peninsula; (2) Anadyr depression (Chukchi Land); (3) Karaginskii Island (northeastern Kamchatka); (4) Hokkaido (northern Japan).

mainly the diatom frustules, represents about 10% of components participating in sedimentation. Accordingly, a high content of biogenic silica is a distinctive feature of bottom sediments in the Chukchi Sea in contrast to those in the other Arctic seas (Pavlidis and Shcherbakov, 2000). In addition to oceanic circulation and sedimentation, the Bering Strait greatly affects the biota development in northern latitudes by promoting exchange of marine organisms between the Pacific and Arctic oceans. The asymmetrical, predominantly northward direction of the biota migrations is noteworthy therewith. Thus, the present-day Bering Strait can be reasonably considered as a typical seaway between two oceans, like other passages connecting the Pacific Ocean with different sea basins.

Geological History of the Strait

The investigation history of the Bering Strait is as long as several decades. The known results show that Eurasia and North America were united by the Bering Land Bridge into a single supercontinent for about 100 m.y., beginning from the Albian Age of the Middle Cretaceous (Marincovich *et al.*, 1990). The Bering Land Bridge favored the exchange of terrestrial animals and plants between the Old and New World, as it is known from geological records. In addition, of the North Pacific, North Atlantic, and Arctic marine biotas are known to have developed separately during the Paleogene and a part of the Neogene. This suggests that the northernmost Pacific represented a giant sea gulf, and essential changes in communities of marine and terrestrial organisms took place in the northern high latitudes only after the Bering Strait opening, apparently in the Late Neogene. The strait opening favored the exchange between Pacific and Arctic marine faunas and floras for the first time but hampered migrations of Eurasian and North American terrestrial biota. However, the Bering Strait was open not permanently: paleontological data evidence for recurrent closures and open-

ings during the last 4.4 m.y. Data obtained by the mid-1990s suggest three periods of the strait opening within the interval of 4.4–2.0 Ma (early–late Pliocene), which stimulated migrations of mollusks between two oceans (Yu. Gladenkov, 1995; all the dates presented in this work are coordinated with the geochronological scale of Berggren *et al.* 1995). The strait opening and closure periods alternated, as it is evidenced by migrations of Asian and American terrestrial organisms. The strait repeatedly closed in the Quaternary time as well and acquired its present configuration at about 0.13 Ma. However, some problems of the Pacific–Arctic connections are still incompletely clear. Above all, we do not know the precise time and cause of the strait opening and the character and changes in direction of marine biota migrations with time. The results obtained recently can elucidate some of these problems.

MATERIAL AND METHODS

New Data on the Bering Strait First Opening

Stratigraphic position of *Astarte* in Neogene sections of northeastern Asia

Despite all the data obtained in the last 35–40 years, the exact time of the first oceanic connection that greatly affected the development of the Arctic-Boreal biota had not been determined until recently. In the 1960s–1990s, some scientists, mostly Russian and American, tried to solve this problem. In general, the estimated age values ranged from 3.6 to 4.4 Ma though some older dates were also considered possible. As mentioned above, the assessments were based on the first occurrence of Neogene mollusks of Pacific origin in the North European sections (Yu. Gladenkov, 1974; Yu. Gladenkov *et al.*, 1980; and others) and on the stratigraphic position of the molluscan genus *Astarte* in the circum-Pacific areas of Kamchatka, Japan, and Alaska (Yu. Gladenkov, 1978). It was assumed that prior to the Bering Strait opening, Cenozoic *Astarte* was absent in the North Pacific but present in the North Atlantic and Arctic regions inhabited in general by biota of North Atlantic origin. *Astarte* penetrated into the Pacific Ocean only after the Bering Strait opening. So, the earliest occurrence level of *Astarte* in the North Pacific sections was expected to be indicative of the strait first opening in the Neogene (Durham and MacNeil, 1967; Yu. Gladenkov, 1976, 1978, 1995; Ogasawara, 1986; Vermeij, 1991). However, the appearance of the oldest *Astarte* in the North Pacific region failed to be precisely dated, because it was impossible to date the event based on the age-diagnostic planktonic microfossils. This issue is disputable even at present.

One of the oldest dates for the Bering Strait opening (4.4 Ma) was established in the Neogene section of the Karaginskii Island, northeastern Kamchatka (Fig. 1), where the *Astarte* appearance level in the upper part of the Limimtevayam Formation (the base of Unit 12) was dated based on diatoms and magnetostratigraphic records (Yu. Gladenkov *et al.*, 1991a, 1991b; *Detailed...*, 1992). The specimens found were ascribed to a new species *A. limimtensis* (*Detailed...*, 1992). However, the date of 4.4 Ma is now doubted, and the corresponding level may be younger. The matter is that the level coincides with the boundary between the *Neodenticula kamtschatica* and *Neodenticula koizumii–Neodenticula kamtschatica* diatom zones (*Detailed...*, 1992), which is defined by the first occurrence of *Neodenticula koizumii* Akiba et Yanagisawa (Fig. 2). According to magnetostratigraphic records, that boundary is inside the interval of reversed magnetic polarity in the Limimtevayam Formation section. This part of the section is interpreted to correspond (*Detailed...*, 1992) to the C3n.1r paleomagnetic episode (the Gilbert Chron) at 4.48–4.29 Ma (Berggren *et al.*, 1995). However, the direct correlation with the paleomagnetic scale indicates that the first occurrence level of *Neodenticula koizumii* in the subarctic Pacific borehole sections is not older than 4.1–4.0 Ma (Barron and A. Gladenkov, 1995). Therefore it is possible that the *Astarte* appearance time in the Karaginskii section may postdate that previously suggested and can be correlative with the C2Ar episode of reversed polarity between 4.18 and 3.58 Ma.

In addition, no attention is paid to the fact that the stratigraphically lowest occurrence of *Astarte* (*Astarte*) sp. is near the base of Unit 10 in the basal part of the Limimtevayam Formation (Yu. Gladenkov, 1972), i.e., about 160 m below the level of first *Astarte limimtensis* (Fig. 3). It is difficult to precisely date the occurrence of *Astarte* sp. because of absence of microfossils at this level. A diatom assemblage found in the underlying Yununvayam Formation (about 140 m below the occurrence level of *Astarte* sp.) is correlative with diatoms of Subzone "a" in the North Pacific *Neodenticula kamtschatica* Zone of Barron and A. Gladenkov (1995) (Fig. 2), which ranges from 7.4 to 5.5 Ma. Stratigraphically higher beds of the Limimtevayam Formation, which are about 80 m above the *Astarte* sp. occurrence level, contain diatoms characteristic of the upper part of the *N. kamtschatica* Zone, most likely of Subzone "c" between 4.8 and 4.1 Ma. Thus, the oldest possible age of earliest *Astarte* may be approaching 7 Ma. However, in the Karaginskii Island section approximately 20 m below the beds with the oldest *Astarte*, there is a hiatus that separates the Limimtevayam and underlying Yununvayam formations. Diatoms have not been found in 120-m-thick deposits below the hiatus. This interval may correspond to a larger part of Subzone "a" of the *N. kamtschatica* Zone. Therefore, the first occurrence level of *Astarte* is most likely younger than 7 Ma. Moreover, the beds situated approximately 35 m higher than the occurrence level of *Astarte* sp. (Fig. 3) contain mollusks *Fortipecten kenyoshiensis* (*Detailed...*, 1992), which first appeared in the earliest Pliocene of the western Kamchatka, Sakhalin, and Japan (Yu. Gladenkov, 1988 and others). Thus, these facts suggest that

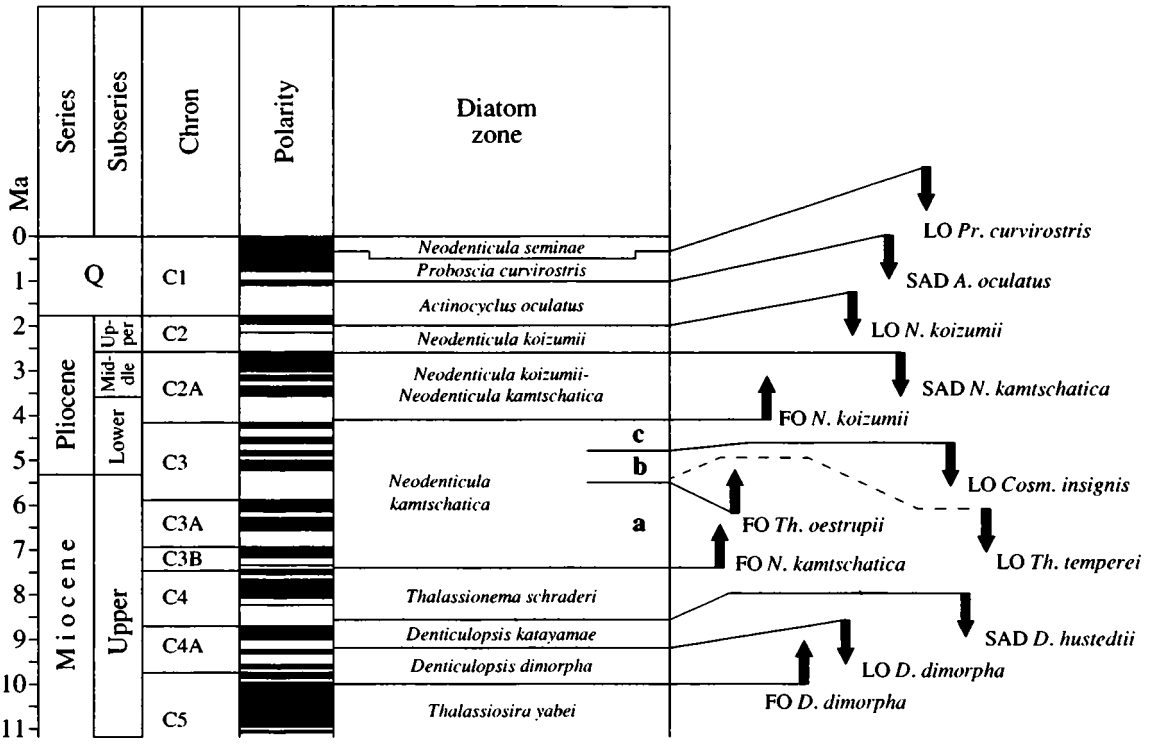

Fig. 2. The North Pacific diatom zonation for the last 11 m.y. (Barron and A. Gladenkov, 1995) correlated with the geochronological and paleomagnetic polarity scales (Berggren *et al.*, 1995); solid lines with arrows indicate datum planes between zones and dotted line denotes the last occurrence level of *Thalassiosira temperei*: (FO) first occurrence; (LO) last occurrence; (SAD) sharp abundance decrease; (a–c) subzones; (*D.*) *Denticulopsis*; (*N.*) *Neodenticula*; (*Th.*) *Thalassiosira*; (*Cosm.*) *Cosmiodiscus*; (*A.*) *Actinocyclus*; (*Pr.*) *Proboscia*.

the first occurrence of *Astarte* in the Karaginskii Island section is restricted by the interval corresponding to Subzone "b" (5.5–4.8 Ma) and probably the upper part of Subzone "a" of the *Neodenticula kamtschatica* Zone, i.e., to the uppermost Miocene–lowermost Pliocene (Figs. 3, 4).

Astarte forms were also found in other regions of northeastern Asia, such as the Chukchi Peninsula (Anadyr depression), Sakhalin, and Japan. According to published data, the oldest *Astarte* occur in the Echinka Formation (270–530 m thick in total) that belongs to the upper part of the Telekai Horizon of the Anadyr depression (Agapitov *et al.*, 1976; Agapitov, 1991) (Figs. 1, 4). *Astarte* cf. (aff.) *montagui* is among the mollusks found in this formation. It is difficult to precisely date that formation lacking planktonic microfossils of biostratigraphic significance, although it may be the latest Miocene in age according to the described impoverished assemblage of benthic foraminifers (Agapitov, 1991).

In Japan, the oldest *Astarte* is known from the Atsuga Formation (about 2000 m thick) of central Hokkaido (Uozumi *et al.*, 1986; Suzuki and Akamatsu, 1994; Ogasawara, 1998; see Fig. 1). The Takikawa–Honbetsu molluscan fauna from the lower part of this formation includes *Astarte (Tridonta) alaskensis*, *A. borealis*, and *A.* sp. (Uozumi *et al.*, 1986; Suzuki and Akamatsu, 1994). The beds with *Astarte* yielded no microfossils, but the fission-track age of tuff beds present 250 m higher in the section is 5.1 ± 0.2 Ma, and the lower age limit of the *Astarte* fauna was previously estimated to be 6.0 Ma. In the early 1990s, diatoms were found in four samples from separate levels of the Atsuga Formation (Sagayama *et al.*, 1992). Their assemblage correlative with that of Subzone "a" of the *Neodenticula kamtschatica* Zone suggests the latest Miocene age of the enclosing deposits. However, these diatoms do not associate with *Astarte* in the section.

In Sakhalin, *Astarte* shells occur in Pliocene deposits (the Pomyr Formation) but their lowermost occurrences have not been precisely dated (Zhidkova *et al.*, 1974). Thus, the available data on the oldest known *Astarte* from Kamchatka, Chukchi Peninsula, and Japan characterize their age as ranging from the latest Miocene to earliest Pliocene (Fig. 4).

Stratigraphic position of *Astarte* in the Pacific coast sections of North America

Until very recently, *Astarte* of late Miocene–early Pliocene age have not been known in North America, although some probably middle Pliocene *Astarte* were found in the Beringian Beds and elsewhere. Additional age information has been obtained recently for the old-

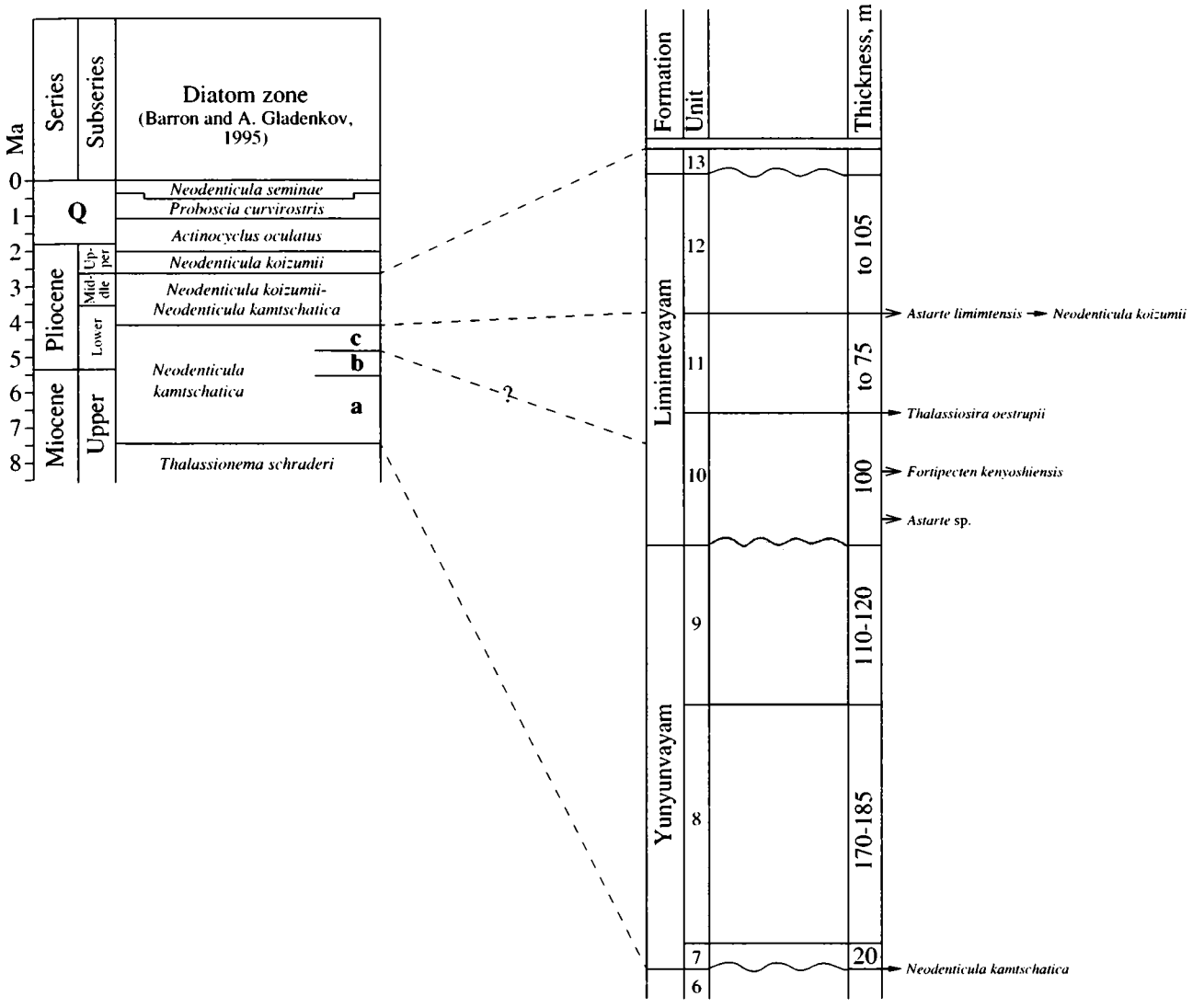

Fig. 3. The upper Miocene–Pliocene section of the Karaginskii Island with appearance levels of molluscan (white arrows) and diatom (black arrows) index species (*Detailed...*, 1992; Yu. Gladenkov, 1972); possible correlation between these levels and the North Pacific diatom zonation is shown (shaded intervals are lacking diatoms; other symbols as in Fig. 2).

est *Astarte* discovered in the upper part of the Bear Lake Formation of the Sandy Ridge section, southwestern Alaska (Marincovich and A. Gladenkov, 1999, 2001; Marincovich, 2000; and others, see Fig. 1). In the 1970s, geologists of the Mobil Oil Company collected there molluscan specimens, which are now stored at the California Academy of Sciences, San Francisco, USA. Reexamination of the collection revealed presence of *Astarte* forms, the oldest of which, e.g., *A. borealis*, are established in two neighboring stratigraphic levels of the section. Shells and casts of fossil mollusks were specially treated to extract the diatom remains. Frequently it is difficult to detect diatoms in the rock matrix because of their frustules disintegration under influence of mechanical and chemical factors during sedimentation, fossilization, and catagenesis. On the other hand, diatom valves can be preserved in shells and concretions buried in fine sediment, being protected from influence of above factors (Barron and Mahood, 1993; and others). This approach was used to discover the Neogene marine diatoms. More than 100 good and satisfactory preserved diatom taxa were found in five molluscan shells from two oldest *Astarte*-bearing levels (A. Gladenkov, 1999, 2001a; and others). In taxonomic composition, the established diatom assemblage appears to be characteristic of Subzone "b" in the North Pacific *Neodenticula kamtschatica* Zone of Barron and A. Gladenkov (1995). The lower boundary of the zone is defined by the appearance of *Thalassiosira oestrupii* (5.5 Ma) and the upper one by the last occurrence of *Cosmiodiscus insignis* (4.8 Ma) (Fig. 2). Both species and zonal index species are components of the Alaskan diatom assemblages. Accordingly, the *Astarte* levels mentioned above are concluded to be the

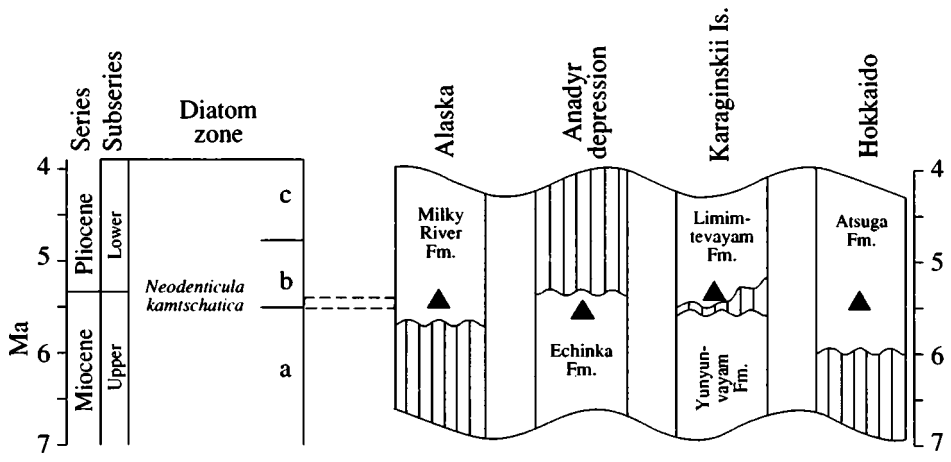

Fig. 4. Stratigraphic position of the oldest Neogene *Astarte* (triangles) in structures surrounding the North Pacific (letter symbols a–c denote subzones)

oldest in Alaska, ranging in age from 5.5 to 4.8 Ma, and the first opening of the Bering Strait was supposed to be within this time interval (Marincovich and A. Gladenkov, 1999, 2001; and others). However, the precise positions of earliest *Astarte* and diatoms in the Sandy Ridge section was established in 1998–1999, when a group of geologists headed by L. Marincovich Jr. restudied the section and sampled the bed-by-bed collection of mollusks and rocks for diatom analysis (A. Gladenkov, 2001b, 2003; Marincovich *et al.*, 2002; A. Gladenkov *et al.*, 2002). The field work showed that the lowermost *Astarte*-bearing interval is in the lower part of the Milky River Formation (Fig. 3) discriminated by Galloway (1974) but not to the upper part of the Bear Lake Formation as supposed previously (Determan *et al.*, 1996). The angular unconformity that separates these formations in the Sandy Ridge section is 28 m below the lowermost occurrence of *Astarte* (*A. borealis*). An assemblage of well-preserved and diverse diatoms was discovered in samples from the level with *Astarte*. The assemblage contains all the forms found there earlier (Marincovich and A. Gladenkov, 1999, 2001) and confirms the previous age interpretation. In addition, the assemblage includes some other age-diagnostic species, the most important of which is *Thalassiosira temperei* (Brun) Akiba et Yanagisawa. Its last occurrence level is a useful Neogene biostratigraphic marker in the North Pacific (Barron, 1980, 1992; Akiba, 1986; Akiba and Yanagisawa, 1986; Yanagisawa, 1990; and others). According to the latest data, this level is dated at 5.4 Ma (Yanagisawa and Akiba, 1998) (Fig. 2). Respectively, the joint occurrence of *T. temperei, T. oestrupii, Neodenticula kamtschatica*, and *Cosmiodiscus insignis* at the same stratigraphic level suggests that the age of the oldest *Astarte* in the Milky River Formation is between 5.5 and 5.4 Ma (Fig. 4). Accordingly, the Bering Strait opened for the first time most likely in the latest Miocene 5.5–5.4 m.y. ago. This conclusion is consistent in general with paleontological data from Kamchatka, Chukchi Peninsula, and Japan, where the oldest known *Astarte* are of a similar age and can be attributed to the latest Miocene. However, possible errors in dating may extend the *Astarte* age range up to the earliest Pliocene inclusive (Fig. 4).

RESULTS AND DISCUSSION

Marine Biota Migrations between Oceans

Impossibility of marine biota migrations through the Bering Strait before the latest Miocene is suggested by data on distribution of Pacific mollusks in the North Atlantic sections. Occurrence of these mollusks is recorded beginning from the Pliocene *Serripes groenlandicus* Zone (with *Clinocardium ciliatum, Macoma calcarea, Modiolus modiolus, Musculus niger, Mya pseudoarenaria, Neptunea decemcostata*, and other forms) in England, the Netherlands, and Iceland, where the zone was established (Yu. Gladenkov, 1976, 1978, 1988; Yu. Gladenkov *et al.*, 1980; and others). In particular, the Sr isotopic age of 4.8–4.7 Ma is determined for the first appearance level of Pacific molluscan species *Mytilus edulus, Modiolus modiolus, Mya truncata, Buccinum undatum*, and *Neptunea angulata*, in the Netherlands (Meijer, 1993), although *Mya* and *Buccinum* forms were previously considered as the late Miocene migrants to the North Atlantic (Strauch, 1970).

On the other hand, there is a serious evidence for invasion of the diatom taxa, which are not older than in the latest Miocene–earliest Pliocene, from the North Pacific into the North Atlantic and Arctic (or vice versa). An example is *Neodenticula kamtschatica*, a typical Neogene species of the North Pacific. It appeared at about 7 Ma and became widespread in the Late Miocene and Pliocene. Being of the Pacific origin, the species is an endemic of the high and middle northern latitudes. Outside the North Pacific, it is known

only from Pliocene deposits of the Norwegian Sea (Dzinoridze *et al.*, 1978; *The History...*, 1979). The invasion of Pacific floral elements through the Bering Strait into the Arctic basin during the time span of 5.5–4.8 Ma is also inferable from the diatom distribution in a section that was sampled by vibradrilling in the Chukchi Sea, the site coordinates 70°55′8″N and 165°57′6″W, water depth 44 m (Barron, private communication of 2001). The assemblage from this site is correlative with Subzone "b" of the *Neodenticula kamtschatica* Zone and includes *Pyxidicula zabelinae, Cosmiodiscus insignis, Thalassiosira gravida, T. oestrupii, Th. jouseae, Porosira* aff. *punctata*, and others (Barron, unpublished data).

It should be noted that the new date of the Bering Strait first opening in the Cenozoic is inconsistent with some previous conclusions based on Arctic diatom assemblages (Zyryanov *et al.*, 1992; Polyakova, 1997). Unlike the high-latitude North Atlantic and North Pacific regions, the vast Arctic region between them is inadequately studied so far for a detailed reconstruction of Cenozoic geological history, paleoceanological changes, climatic events, and marine biota development. This is mostly a consequence of difficulties with deep-sea drilling in the Arctic Ocean covered with ice. The reliable stratigraphic data were unknown for this region, and the Arctic Ocean was like a peculiar blank spot until the recent time. However, the Arctic basin had, like now, a great influence on the Earth climate (cold water masses and the ice cover). Being connected with the North Pacific and North Atlantic basins, it determines the biota development in these oceans as well. Polyakova (1997) tried to reconstruct the Late Cenozoic history of the Arctic seas based on diatom floras known from onshore sequences and reported her conclusions on water exchange between the Arctic and the North Pacific oceans in the Cenozoic and on time of the Bering Strait first opening.

Polyakova, who analyzed the Neogene marine diatoms from lowlands of the northern Yakutia and northern Chukchi Peninsula, inferred their North Pacific origin based on taxonomic similarity between the Arctic and Pacific assemblages. The Miocene and Pliocene assemblages defined by means of correlation with the North Pacific diatom zonation allowed Polyakova to conclude that of the Arctic and Pacific oceans became connected in the Miocene. In her opinion, the Arctic biota developed under influence of the Pacific biota since the Middle Miocene, and the Arctic and Pacific oceans were connected, most likely via the Bering Strait region, in the terminal early Miocene–initial middle Miocene and periodically afterward, beginning from the terminal middle Miocene to the present time (Polyakova, 1997). However, these conclusions, which are based on a single fossil group studied in incomplete sections, seem disputable and doubtful. First, all stratigraphically important taxa Arctic diatoms listed by Polyakova are widely distributed in Subboreal and Arctic-Boreal regions, because they were recovered by the deep-sea drilling not only in the North Pacific but also in the Norwegian–Greenland basin and the Labrador Sea of the North Atlantic (A. Gladenkov, 2001a; Marincovich and A. Gladenkov, 2001). Accordingly, these species could migrate into the Arctic from the North Atlantic basins rather than from the Pacific ones. Moreover, the early–middle Miocene taxa of unambiguously North Pacific origin have not been found in the Arctic assemblages. In the middle to high latitudes of the North Pacific, successive datum levels of Neogene diatoms are identified in relatively complete sections and calibrated with magnetostratigraphic records. In contrast, stratigraphic succession of scarce diatom assemblages from the Arctic lowlands has not been similarly substantiated and controlled by the other fossil groups. In addition, fossil diatoms, as is known, can be transported by wind from burial places over considerable distances (Burckle *et al.*, 1988; Kellogg and Kellogg, 1987, 1996; and others). Therefore, we do not exclude a possibility of wind transportation for Miocene diatoms, which could be brought from the North Pacific into the Arctic and subsequently reworked. In view of this possibility, Polyakova's supposition should be regarded with a great caution.

Possible Causes of the Bering Strait Opening in the Late Miocene

As noted above, recent data on Neogene mollusks and diatoms suggest the first Cenozoic opening of the Bering Strait about 5.5–5.4 Ma ago. Analysis of available geological materials reveals two main causes of this event.

First, the strait opening in response to tectonic processes can be inferred from summarized data on tectonic structure of the Bering Sea–Aleutian region (Scholl *et al.*, 1992; Plafker and Berg, 1994; Mackey *et al.*, 1997; Imaev *et al.*, 2000; and others). According to suggested reconstructions, tectonic processes in eastern and northern parts of the Bering Sea–Aleutian region might induce the crustal extension in the Bering Strait area, which accelerated 6.0 Ma ago or later. It is possible that the acceleration caused by intense tectogenesis movements resulted in crustal thinning and subsidence of the crustal blocks thus giving rise to the Bering Strait formation. The modern structure of Kamchatka began to develop under influence of the tectonic activation in the second half of the late Miocene. The intense faulting and folding were characteristic of the Kamchatka region in the latest Miocene (6.0–5.5 Ma ago). In particular, large tectonic nappes were formed at that time in East Kamchatka (Yu. Gladenkov and Shantser, 1989, 1990).

Thus, the tectonic activation of the latest Miocene took place simultaneously in different parts of the Bering Sea–Aleutian region. The crustal thinning or block subsidence in response to tectonic processes in the land area that connected Asia and North America could bring about a relative sea-level rise, and both factors

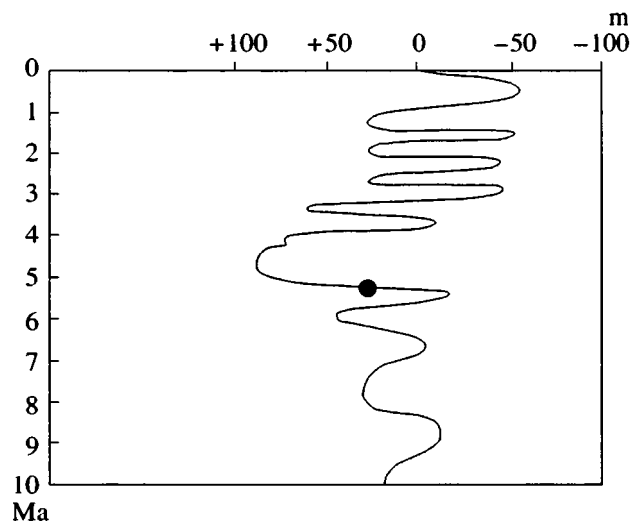

Fig. 5. Eustatic sea level fluctuations (Haq *et al.*, 1987) during the last 10 m.y. correlated with the geochronological scale (Berggren *et al.*, 1995); black dot denotes the supposed first opening of the Bering Strait in the Neogene as inferred from diatom data.

could together breach the Bering Strait after formation of a tectonic step in that area 6.0 Ma ago. It is apparent that the age estimated for the regional tectonic activation is well consistent with fossil records suggesting the initial strait opening approximately 5.5–5.4 Ma ago.

Second, the Bering Strait opening within the above time interval might be related to a global eustatic sea-level rise. The largest transgression in the last 12 m.y. took place 5.0 Ma ago, when sea level reached a peak value of 70 m above the modern level (Haq *et al.*, 1987; Fig. 5). That transgression started however earlier in the warming period that postdates the Messinian salinity crisis (6.0 Ma) and subsequent sea-level drop caused by cooling about 5.8 Ma ago (Haq *et al.*, 1987; Fig. 5).

It is remarkable that the Bering Strait opening was close in time to the Mediterranean transgression after the Messinian crisis. A detailed oxygen isotope analysis of the Neogene sections in the equatorial East Pacific (ODP Site 846) and northwestern Morocco (Salé Borehole) revealed a distinct trend of $\delta^{18}O$ decreasing, with rapid and extreme fluctuations, during the initial Subhron C3r of the Gilbert Polarity Zone (Shackleton *et al.*, 1995; Hodell *et al.*, 1994). The lowest $\delta^{18}O$ values correspond to the global sea-level rise during the warmest interglacial, which could be responsible for the Mediterranean flooding after the salinity crisis. The event occurred within the TG9 isotopic stage (5.46 Ma) according to one interpretation (Shackleton *et al.*, 1995) and across the TG12–TG11 boundary (5.32 Ma) according to the other one (Hodell *et al.*, 1994). This indicates that the Mediterranean transgression took place either in the latest Miocene or across the Miocene–Pliocene boundary.

Of course, the Mediterranean transgression and the Bering Strait opening could be caused by different factors responsible for disproportional sea-level rise in remote regions. Anyway, the rapid and extreme eustatic fluctuations (Shackleton *et al.*, 1995; Hodell *et al.*, 1994) approximately coincide in time with the Bering Strait opening. The date of 5.46 Ma established for the minimal $\delta^{18}O$ values (Shackleton *et al.*, 1995) is consistent with the time of the Bering Strait opening (5.5–5.4 Ma) suggested based on new paleontological data. On the other hand, as noted above, the upper age limit of 5.4 Ma for the strait opening is defined based on the index diatom species *Thalassiosira temperei* found in the lower part of the Milky River Formation of Alaska. The last occurrence level of this form in the North Pacific is dated by means of sedimentation rate extrapolation in deep-sea drilling sections. It is possible to assume therefore that *Thalassiosira temperei* disappears slightly earlier at about 5.3 Ma in the Alaska section. Correspondingly, we do not exclude that the Bering Strait opening could take place at 5.32 Ma (Hodell *et al.*, 1994), i.e., at the time of the Miocene–Pliocene boundary.

Noteworthy is the increasing abundance of warm-water taxa in the latest Miocene diatom assemblages of the North Pacific. This evidences for climatic warming at that time. The presence of some warm-water species in the diatom assemblages of the Gulf of Alaska and the Milky River Formation may be also as a response to influence of relatively warm oceanic water (A. Gladenkov, 2003). In sections of Alaska, Kamchatka, and Hokkaido, just below the lowermost *Astarte*-bearing stratigraphic level, there are hiatuses or unconformities and lithological changes indicative of a marine transgression. Consequently, the Bering Strait opening might be caused by flooding of land in that region during the latest Miocene marine transgression and significant sea-level rise.

Thus, the Bering Strait first opening at 5.5–5.4 Ma might be a consequence of either tectonic events or eustatic sea-level rise. A combined influence of both factors cannot be excluded as well.

Possible Changes in Oceanic Circulation and Marine Biota Migrations through the Bering Strait in the Early Pliocene

According to the latest data, the deep-to-intermediate water exchange between the Pacific Ocean and Caribbean Sea via the Panama Strait ceased at 4.6 Ma, because the latter turned out to be less than 100 m deep (Haug and Tiedemann, 1998). The event significantly affected oceanic circulation, state of deep water, and climate in the Pacific and Atlantic oceans. The simulated global oceanic circulation shows that prior to this event the dominant flow through the Bering Strait could be oriented southward, from the Arctic Ocean (influenced by the North Atlantic water inflow) into the North Pacific (Maier-Reimer *et al.*, 1990), i.e., opposite

to the present-day direction. After rearrangement of the global oceanic circulation in response to the Panama Strait shoaling, stream through the Bering Strait should be oriented northward, i.e., from the North Pacific into the Arctic and farther into the North Atlantic. This suggests an increased supply of fresh cold water to the high-latitude North Atlantic. Thus, the shoaling (and subsequent closure) of the Panama Isthmus and the Bering Strait opening were significant Neogene events that brought about important hydrological and paleogeographic changes. The events had not only regional but also global effect, as they connected the Pacific and Atlantic oceans through the Arctic and resulted in transformation of oceanic circulation and climate in the Northern Hemisphere. (Maier-Reimer *et al.*, 1990; Reason and Power, 1994; Shaffer and Bendtsen, 1994; and others). All the factors must influence the marine biota development and migration ways of different organisms. According to Marincovich (2000), taxonomic composition of the Neogene high-latitude molluscan assemblages is consistent with the Pliocene reversal of marine flow through the Bering Strait. During the latest Miocene–earliest Pliocene (prior to the closure of the deep Central American seaway), mollusks predominantly migrated through the opened Bering Strait from the North Atlantic and Arctic into the North Pacific. The reversed migrations after 4.6 Ma are evidenced by appearance of the North Pacific species in the North Atlantic beginning from 3.6 Ma. However, migrations of marine benthic fauna might be more complicated. Mollusks of the Pacific origin were found in the Netherlands borehole sections at the level of 4.8 to 4.7 Ma (Meijer, 1993), while mollusks of the North Atlantic origin were discovered in the North Pacific sections at the level of 4.6 Ma and higher (Yu. Gladenkov, 1978, 1988, 1995; and others).

We should add one more consideration concerning a problem, which has not been discussed yet. Let us suppose that prochoreses in both the Pacific and Arctic oceans were simultaneous despite the asymmetrical biota migration through the Bering Strait from the Pacific into the Arctic and *vice versa*. This puts forward an important problem of relative migration rates. If the Bering Strait opened about 5.5–5.4 Ma ago and Pacific migrants appeared in the North Atlantic 4.8–4.6 Ma ago (see above), the time span more than 0.5 m.y. (up to 0.9 m.y.) was necessary for mollusks of Pacific origin to cross the Arctic Ocean and spread over its shelf zones. Data from Iceland only suggest the time span of 1.8 m.y. It is unclear therewith whether we deal with real rates of the Pacific mollusks dispersal over the Arctic shelf zones or the assessments are a consequence of incomplete geological records on the migration ways and dispersal of mollusks, which occupied the Arctic basin. These problems can considered now only in provisional aspect, but we should remember them, if we want to define whether the Arctic Ocean was occupied instantaneously or not in terms of geological time.

There is no data implying that the North Pacific diatom assemblages penetrated into high latitudes of the North Atlantic through the Bering Strait before the early Pliocene. However, the earlier interchange between the North Pacific and North Atlantic floras was possible through the Panama seaway, as suggested by the presence of common planktonic elements in the North Pacific and North Atlantic diatom assemblages, although most of their taxa are cosmopolitans. However, some transitional warm-water taxa, such as *Thalassiosira praeoestrupii* and *T. oestrupii*, appeared asynchronously in two oceanic basins.

Climatic Events at the Bering Strait Formation Time

The ice-rafted material was found at separate levels (6.7 to 4.4 Ma) of the ODP core sections penetrated in the northwest Pacific and Gulf of Alaska (Krissek, 1995; Rea *et al.*, 1995). The oldest ice-rafted material in the Neogene Yakataga Formation exposed along the Gulf of Alaska coast is recorded at the level of about 6.2 Ma (Lagoe *et al.*, 1993). Lithology and paleontological remains from this formation imply the tidewater glaciations in the Gulf of Alaska between 6.2 and 4.4 Ma during the climate deterioration in the high northern latitudes (Lagoe *et al.*, 1993). "Floating" ice-rafted debris are also characteristic of the coeval interval of the Yununvayam and Limimtevayam formations on the Karaginskii Island, northeastern Kamchatka (*Detailed...*, 1992). In combination with paleontological data, this evidences for a cooling and water temperature drop in the Kamchatka region. Noteworthy is scarcity of ice-rafted material in the ODP cores within the interval of 6.7–4.4 Ma. Moreover, abundance of this material usually decreases away from the coast and southward. Hence, coasts of Alaska and Kamchatka could represent at that time areas of local glaciations and the ice-rafted debris transportation to the ocean (Krissek, 1995). The supposition is consistent with composition of the Neogene diatoms from the Milky River Formation, which were deposited on the Alaska shelf. The latest Miocene–earliest Pliocene diatom assemblages contain here some marine Arctic-Boreal species (*Bacterosira fragilis, Detonula confervacea, Porosira glacialis, Thalassiosira kryophila, Th. gravida*, and others) and the forms that commonly live near sea ice fields (*Fragilariopsis oceanica, Fr. cylindrus* and others), the vegetation periods of which were likely coincident with ice melting near the southwestern margin of Alaska (A. Gladenkov, 2003). On the other hand, presence of planktonic taxa characteristic of the subtropical and temperate regions (*Thalassiosira leptopus, Th. temperei, Th. oestrupii, Auliscus* sp., *Bacteriasrum varians*, and others) indicates influence of relatively warm oceanic surface water.

In general, the latest Miocene–earliest Pliocene diatom assemblages of the Subarctic Pacific contain an increased amount of warm-water forms, which suggests a relatively warm period (Oreshkina, 1986; and

others). In the early Pliocene, after 5.1–5.0 Ma, cold-water taxa increased in abundance and some new species (*Thalassiosira latimarginata* and *T. tertiaria*) appeared. In the period of the second half of the early Pliocene to the initial middle Pliocene, warm-water diatom species intensely penetrated into high latitudes (Oreshkina, 1986; Barron, 1995; and others). Warm-water benthic taxa also appeared at that time in the Subarctic circum-Pacific areas (*Detailed...*, 1992; and others). The absence or reduced amount of ice-rafted material in the high-latitude North Pacific, the Gulf of Alaska included, also implies the climatic warming about 4.6–3.1 Ma ago (Lagoe *et al.*, 1993; Krissek, 1995; Rea *et al.*, 1995).

Diatom Productivity in North Pacific during the Late Miocene–Pliocene as Related to Climatic and Oceanographic Changes

In the North Pacific, the time span from about 6.4 to 2.7 Ma was characterized by increasing biogenic silica mass accumulation (BSMA) and by the maximum diatom productivity during the Cenozoic (Rea *et al.*, 1993, 1995; Rea and Snoeckx, 1995; Barron, 1998). The enhanced biogenic silica deposition is recorded in the marginal basins (Sea of Okhotsk, Bering Sea, California Gulf) and in western and eastern sectors of the Subarctic Pacific. In the western sector, where the sediments are mostly represented by siliceous ooze, sedimentation rates were 4.5 times greater than before beginning from 6.4 Ma and remained high until 2.7 Ma ago with their maximum of 14 to 15 cm/ka recorded between 4.6 and 3.1 Ma (Rea *et al.*, 1993, 1995). The BSMA values increased from about 0.7 to more than 3.0 g/(cm^2 ka) and then, after 2.7 Ma, it sharply decreased down to the values recorded prior to 6.4 Ma. A similar trend was characteristic of the Gulf of Alaska, where after 6.4 Ma sedimentation rates increased sharply (approximately four times), reached the maximum values of more than 4 cm/ka, and the BSMA grew from about 0.6 to 1.5–2.0 g/(cm^2 ka). In the interval from about 3.1 to 2.7 Ma, sedimentation rates declined from 3 to 1.5 cm/ka and the BSMA values from ~1.0 to 0.2–0.3 g/(cm^2 ka).

According to these data, the peak of biogenic silica sedimentation and diatom productivity in the subarctic Pacific (4.6–3.1 Ma) coincides in time with a significant warming. The beginning of above period (4.6 Ma) is correlative with the Panama Strait shoaling and cessation of deep-to-intermediate water exchange between the Pacific Ocean and the Caribbean Sea (Haug and Tiedemann, 1998). The latter event should accelerate inflow of "ancient" Antarctic deep-water masses rich in nutrients into the North Pacific because of reorganized oceanic circulation. On the other hand, the Pliocene warming could trigger inflow of large masses of relatively warm saline surface water into the North Pacific from the south (Barron, 1995). A wide zone that separated warmer and cooler surface waters was likely favorable for penetration of warmer, more saline surface waters into the high-latitude North Pacific, where they diminished the vertical stratification of water masses and caused water mixing and upwelling of nutrient-rich deep waters (Barron, 1998).

UNSOLVED PROBLEMS

A great progress has been achieved in research of the Bering Strait geological history since the 1960s. Most of the new data are summarized in three publications: "The Bering Land Bridge" (1967), "Beringia in the Cenozoic Era" (1976), and "Beringian Paleoenvironments" (2001). Some problems were attacked in monographs (Strauch, 1972; Yu. Gladenkov, 1978) and in some articles (see above). However, there is a number of problems that have not been adequately studied or solved. Some of them partially mentioned above are outlined below.

(1) Stratigraphic distribution of old astartids in the North Pacific is not very clear so far. They are known from Mesozoic (Zakharov, 1970) and Lower Paleogene deposits of the region, in particular, from the Paleocene Sinegorsk Beds of Sakhalin (Kalishevich *et al.*, 1981). Why are astartids missing from the Eocene and Miocene intervals of the North Pacific sections, except for the uppermost Miocene? There is no well-reasoned answer.

(2) A problem of the first connection between the Arctic and Pacific oceans is still under discussion. The common opinion is that communication between two oceans started with the opening of the Bering Strait. However, prior to the late Miocene–Early Pliocene, they might be connected via an area situated further to the west, in northwestern Asia (northern Yakutia). An ancient strait can be outlined there based on of sporadic finds of marine diatoms (Polyakova, 1997). Although this supposition is not substantiated well, it needs to be studied and checked.

(3) The migration ways of Pacific benthic organisms from the Bering Strait to the North Atlantic, i.e., across the Arctic Ocean, are undetermined yet. It is not clear what shelf zones, Eurasian or North American, were populated predominantly by these organisms. Without relevant data, it is impossible to reconstruct the ancient sea currents and former hydrological regimes, which were essential for migrations of the North Atlantic assemblages, mollusks of the genus *Astarte* included, into the Pacific and the Pacific assemblages with *Mya* and *Neptunea* forms into the North Atlantic. Physical and climatic barriers may be also of great importance. For instance, shoal area of the New Siberian Islands is a good example of a physical barrier, which controls the biota distribution in the Arctic shelf. The specific climatic regime of this area with a low seawater temperature, pack ice sheets, and fresh water influx along them controls the biota distribution. Apparently, the Neogene climatic conditions differed from the Quaternary ones

in higher temperatures of seawater. Otherwise, Boreal mollusks of the Pacific origin could not cross the Arctic-type polar basin. What way (taking into consideration the former sea currents) was used by the North Atlantic migrants to reach the Bering Strait is unclear. Future deep-sea drilling in the Arctic basin may answer this question.

(4) There are diverging assessments of biota migration rates in the shelf zones of oceans under consideration, especially in those, where migrants arrived first after passing the straits from the basin of their origin. In the light of available data, it is unclear whether the migrations were virtually instantaneous or lasted over a period of 0.5–2 m.y.

(5) As stated above, it is difficult to estimate unambiguously the role of tectonic and eustatic factors in the Bering Strait opening. It is likely that both factors were jointly responsible for this event. The sea-level rise was likely brought about by the global warming, and the increased temperature of Arctic surface waters allowed the Boreal migrants of the Pacific origin to cross more than 6000–7000 km of the Arctic shelf and to reach the North Atlantic.

(6) The commonly recognized factors influencing the biota migrations are hydrodynamic regimes, marine currents, climatic changes, and eustatic fluctuations. As a rule, no attention is paid however to some species expansion supplanting the others in biocoenoses. Although such a competition is insufficiently clear and proved, it does exist. Similarly, competition between neighboring ecosystems in the transitional zones may lead to expansion of some ecosystems and to reduction of the others. It is the activity of the North Pacific ecosystem, as well as the relative abundance and diversity of its biotic components, that is considered as responsible for dominating direction of biota migrations from the Pacific into the Arctic (Yu. Gladenkov, 1988; Vermeij, 1991; and others).

(7) The asymmetry of biota migrations through the Bering Strait has been already mentioned. The asymmetry ratio estimated for the Pliocene is 8 : 1, meaning that for every eight species migrated from the Pacific into the Arctic, there was only one species migrated in the opposite direction (Durham and MacNeil, 1967; Yu. Gladenkov, 1974). The ratio variations can be attributed to different factors, such as changes in hydrological and climatic regimes, opening and closing of other straits, and so on (Marincovich, 2000). As it is difficult to define a general trend of these variations with geological time, the problem needs to be investigated further.

(8) We do not fully realize the influence of migrants on biocoenoses development in the Arctic and Pacific ecosystems. Occupying certain ecological niches, migrants may become important components of an ecosystem. For example, astartids play now an important role in the North Pacific assemblages and represent index-species (*Astarte rollandi, A. alaskensis, A. ioani* and others) of modern Boreal biocoenoses. After a certain period, migrants may become intrinsic in biocoenoses and, as such, they may be considered as native elements of assemblages at the next geological stage.

Conversely, some newcomers failed to occupy a niche in new biocoenoses. They become extinct or pushed aside with time. A good example is the peculiar Arctic fauna, which was formed under the Pliocene–Quaternary climatic conditions. Foreign Boreal elements had either to move to the south or to become more resistant to cold environment. As a result, after significant cooling of the Northern Hemisphere 2.7 Ma ago and, especially, in the Quaternary time, first the Arctic-Boreal and then Arctic assemblages appeared. The molluscan assemblage with *Portlandia arctica* appeared in the Arctic basin in the Early Quaternary (the Eopleistocene of Iceland) and spread widely over all shelf zones of the region. During the periods of intense cooling, this assemblage migrated to the North Pacific through the Bering Strait. It was found in the Karaginskii Island stratigraphic section of northeastern Kamchatka.

CONCLUSION

The latest data imply that the Bering Strait formation began in the latest Miocene–earliest Pliocene (5.5–5.4 Ma ago according to diatom records). The strait first opened in the Neogene because of regional tectonic processes or sea-level eustatic fluctuations; a combined effect of these factors is also possible. Since that time the North Pacific ecosystem became intermittently unclosed in its northernmost part and greatly influenced the biota distribution in northern latitudes. The migration direction of marine biota through the Bering Strait was reversed 4.6 Ma ago probably because of the oceanic circulation reorganization induced by the paleoceanographic changes, particularly in the Panama Isthmus region.

The Bering Strait was one of the most important seaways in the Neogene history of the Pacific and Arctic Oceans. It is an interesting object for a methodical geological study. The unsolved problems listed above clearly show how important is to understand the geological history of similar objects and their influence on holarctic and global processes.

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Age of Terrigenous Rocks of Northeastern Karaginskiĭ Island (Eastern Kamchatka)

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Abstract—Terrigenous complexes of the Cenozoic accretionary prism are exposed in the northeast of Karaginskiĭ Island (eastern Kamchatka). Their age is debatable to the present day. The studied nannoplankton from mudstone and the fission-track dating of detrital zircon from sandstone elucidate agè of terrigenous rocks. Aleuropelites yield two nannoplankton assemblages. One ranging in age from the Ypresian–Lutetian (Supra-Khynkhlonai sequence) to Bartonian–Oligocene (sedimentary melange) is considered as an *in situ* assemblage characterizing age of sediments. The other one consisting of Upper Cretaceous–Paleocene taxa is regarded as redeposited from older sediments of unknown provenance. According to fission-track dating, the youngest population of detrital zircons is close in age to the deposition time of the most rocks studied. Terrigenous rocks of the Karaginskiĭ Island tend to be younger southeastward off the axial zone. The flysch complex is dated back to the Bartonian–early Oligocene, whereas rocks of sedimentary melange range in age from the Oligocene to early Miocene according to the fission-track dating of detrital zircon. Younger ages of rock located closer to the trench are typical of accretionary prisms.

Key words: nannoplankton, fission-track dating, detrital zircon, accretionary prism, Karaginskiĭ Island, Kamchatka.

INTRODUCTION

The Karaginskiĭ Island represents a part of the Komandor Basin framing (Fig. 1). The island structure and the history of its formation is important for understanding of geodynamics of the youngest tectonic zones in northeast Asia and adjacent sea areas. According to available reconstructions (Chekhovich *et al.*, 1989; Chekhovich *et al.*, 1990; Sokolov, 1992; Shapiro, 1995; *Ob'yasnitel'naya zapiska...*, 2000; Solov'ev *et al.*, 2002), the continental margin structure has been formed by collision of a Cretaceous island arc with the northeastern Asia. As is supposed, the collision resulted in formation of a new subduction zone east of the island arc accreted to the continent, which consumed first the Pacific Plate and then minor plates of the Komandor Basin of the Bering Sea. The Karaginskiĭ accretionary prism, which is exposed at present in the eastern part of the island, formed above that subduction zone (Chekhovich *et al.*, 1989). The prism should be paragenetic to volcanic belts of the Kamchatka Peninsula and southern Koryak Upland (Bogdanov *et al.*, 1987; *Ob'yasnitel'naya zapiska...*, 2000), i.e., to the Eocene–Oligocene volcanics of the northwestern Govena Peninsula (Govena arc), middle–upper Miocene volcanics of the Penzhina Bay coast (Kinkil belt), and Miocene–Pliocene volcanics of the Median and Vetvei ridges (Central Kamchatka and Apuka–Vyvenka belts). One of the difficulties complicating reconstruction of these relationships is the lack of age data on rocks of the prism. This work presents the results of fission-track dating for detrital zircons from the prism sandstones and new data on nannoplankton from mudstones. Furthermore, age data for the tuffaceous–terrigenous rocks crowning the island-arc volcanics in the axial part of the island are considered in addition.

GEOLOGIC SETTING

Three major structural zones are distinguished in the Karaginskiĭ Island (Chekhovich *et al.*, 1990; Kravchenko-Berezhnoy and Nazimova, 1991). The axial zone is mainly composed of gabbro–ultramafic rocks of the ophiolite association and of the Late Cretaceous– Paleocene island-arc volcanic and sedimentary rocks formed far to the south of their present position according to paleomagnetic data (Kovalenko, 1990; Kovalenko *et al.*, 1999; Kovalenko and Kravchenko-Berezhnoy, 1999). Some researchers (Shapiro, 1995;

Kovalenko *et al.*, 1999) believe that the volcanic rocks, the upper part of which is distinguished as the Khynkhlonai Formation, belonged to an island arc that collided with the continent in the middle Lutetian (Solov'ev *et al.*, 2002). At its terminal Paleogene stage of evolution, the collided structure is known as the Govena arc (Chekhovich *et al.*, 1990).

In the northeastern part of the island, volcanics of the Khynkhlonai Formation are conformably overlain by a mudstone member with tuffite laminae, the age of which has not been exactly established. Some researchers (Shapiro and Petrina, 1984; Chekhovich *et al.*, 1990) attribute these rocks to the flysch complex characterized below. This is not undisputable however, because interlayers of sandstone (plagioclase + quartz + rhyolite) typical of the flysch are missing from this sequence, whereas tuff and tuffite, not characteristic of flysch, are abundant. Furthermore, their conformable occurrence above the island-arc volcanics is inconsistent with the sequence position in the accretionary prism. We separate it therefore from the flysch and designate arbitrarily as the Supra-Khynkhlonai sequence, the age of which determines the upper limit of the Govena arc active development. Previously, the sequence age was determined based on correlation with the flysch complex as ranging from the Maastrichtian–Paleocene to middle Eocene (Shapiro and Petrina, 1985; Chekhovich *et al.*, 1990).

The axial zone is bordered on northwest, mainly through faults, by a monocline composed of the middle Eocene–Pliocene, conformably bedded sediments, which dips in the northwestward direction, being complicated by gentle folds, and represents a slope of the Litke trough. Lower levels of the trough axial section are exposed on the Il'pin Peninsula (Volobueva *et al.*, 1994), where Upper Cretaceous volcanics are overlain with a minor unconformity by tuff–terrigenous rocks of the lower Paleocene. Another unconformity is established in the middle Miocene interval of overlying rocks only (Gladenkov, 1972; Gladenkov *et al.*, 1998). Faults bounding the Litke trough of the Karaginskiĭ Island are of a small amplitude. In places, Miocene and Pliocene sedimentary sequences occur with unconformity on rocks of the island axial zone.

According to Chekhovich *et al.* (1989, 1990), the southeastern part of the island is composed of the flysch and sedimentary melange complexes. *The flysch complex* exhibits the variably ordered alternation of mudstones and sandstones, some of which represent typical turbidites. However, many mudstone members of this complex up to hundreds of meters thick are lacking sandstone interlayers. An important feature of flysch sandstones is prevalence in their composition of acid volcanic and, to a lesser extent, granite clasts over fragments of more basic volcanic rocks (Shapiro, 1984; Shapiro *et al.*, 2000). Characteristic of some mudstone members in the flysch are well-rounded disseminated pebbles of acid volcanic rocks (Shapiro, 1984; Shapiro *et al.*, 2000). Flysch and mudstone members often enclose lenticular intercalations of tholeiitic pillow lavas, which are regarded either as conformable lenses, or as olistoliths and tectonic detached masses of the oceanic crust (Chekhovich *et al.*, 1990; Kravchenko-Berezhnoy *et al.*, 1990). Beds of the flysch complex are intensely folded or form northwestward-dipping monoclines separated by imbricated thrust faults conformable with bedding in general but in places outlined by protrusions and lenses of serpentinite melange. The flysch complex is bordered by major, northwestward dipping reversed or thrust faults, which separate it from volcanics and ophiolites of the axial part.

The intensity of folding of the flysch complex increases in the southeastward direction, where the beds losing their coherence are dismembered into large or small, often rotated lenses. Among the latter, there appear large lenses of pebbly mudstones with disseminated pebbles and boulders of rhyolites and their tuffs. The transition to the *sedimentary melange complex* is marked by distinct lithological changes: sandstones are less frequent in the section, being replaced by siliceous mudstones, cherts, tuffites and tuffs, psephitic included. A kind of chaotic mixture of nearly isometric fragments of different rocks set in clay matrix is often characteristic of exposures, though coherent members of tuffs and cherts can also be observed. In places, lenses of pillow or massive (rarer) basalts are confined to tuff interlayers, while small basalt bodies often occur in chaotic members. In terms of geochemistry, basalts are frequently of the oceanic type, though the island-arc rocks also associate with them (Chekhovich *et al.*, 1990). Like in the flysch complex, rhyolite pebbles prevail among rock debris in pebbly mudstones. The thrust fault separating flysch from sedimentary melange complexes is traceable northeastward almost up to the shore near the Nizkii Cape, where it turns southwards, extending for 7 km and outlining antiform hinge, in the eastern flank of which the overturned basal planes of sandstone and tuff beds are facing westward (Fig. 1).

The age interpretation of rock complexes in the northeast of the Karaginskiĭ Island changed with time. In the first publication (Kharkevich, 1941), they were referred to as Mesozoic. Later on, the Miocene molluscan fauna was found in rocks of both complexes exposed along the southeastern coast of the island (Khramov *et al.*, 1969). In 1967, G.P. Borzunova who sampled sedimentary melange once more suggested the Oligocene age of its fauna. In opinion of Yu. B. Gladenkov (Chekhovich *et al.*, 1990, and private communication), the described mollusks are of the middle Eocene age. The Eocene age of mollusks collected from the flysch complex was also determined by V.M. Gladikova (Mel'nikova and Dolmatov, 1973). Afterward, agglutinated foraminifers from flysch and sedimentary melange complexes were referred to the Maastrichtian–Danian, similar to their assemblages from the Vetlov Formation of the Kumroch Ridge (Shapiro and Petrina, 1985). At present, the formation is attributed to the

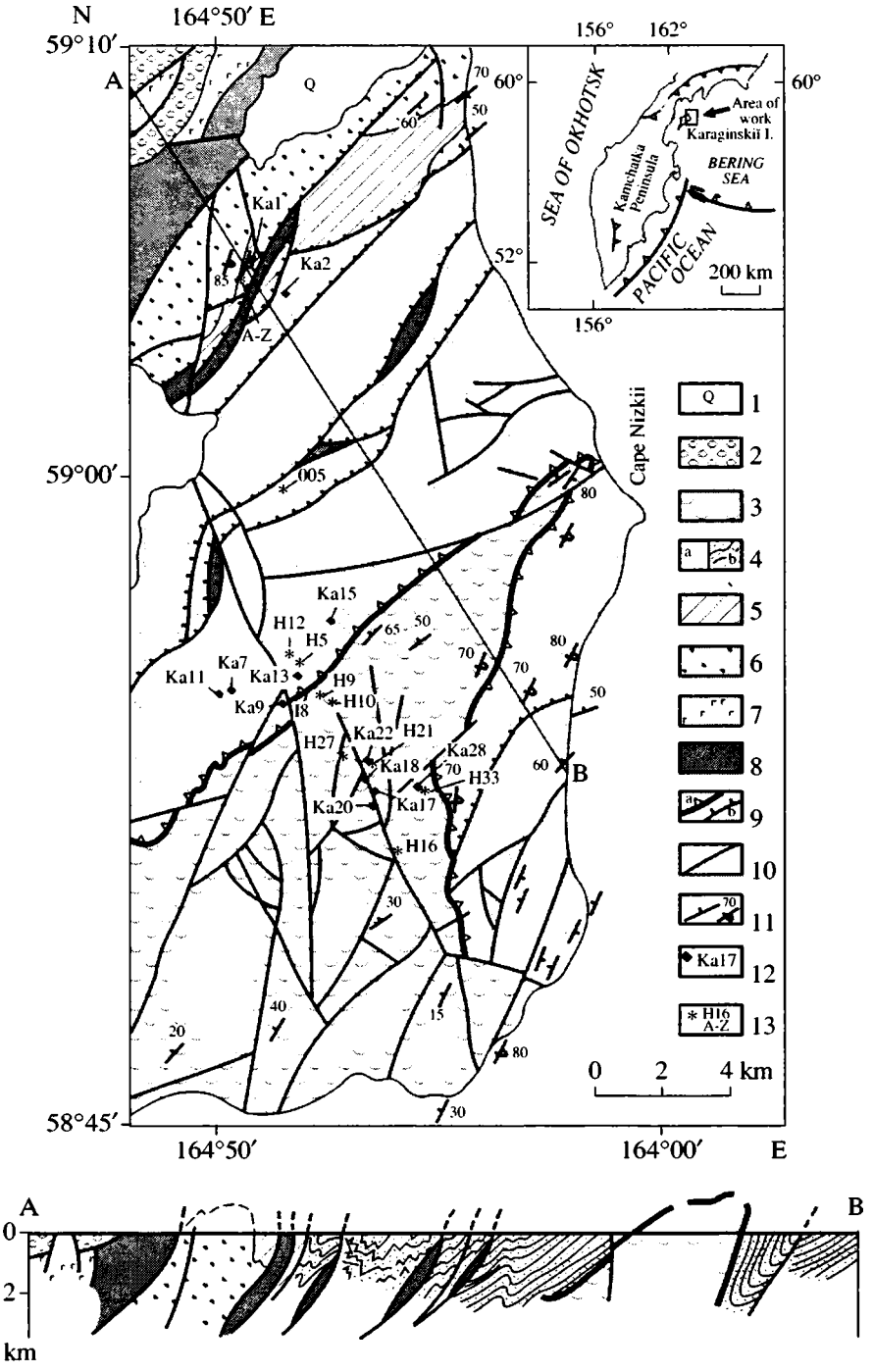

Fig. 1. Geological structure of the northeastern Karaginskiĭ Island and geological profile along line AB (modified after Shapiro and Petrina, 1985): (1) Quaternary deposits; (2) Neogene deposits; (3) sedimentary melange complex (Oligocene–Lower Miocene); (4) flysch complex (middle Eocene–Lower Oligocene) on scheme (a) and in profile (b); (5) Supra-Khynkhlonai sequence (Lower Eocene); (6) Khynkhlonai Formation of basaltic andesite and tuffs (Maastrichtian–Lower Eocene); (7) pre-Maastrichtian Cretaceous deposits; (8) gabbro-ultramafic rocks of ophiolite association; (9) thrust and reversed faults connected with flysch complex thrust over sedimentary melange complex (a) and others (b); (10) steeply dipping faults; (11) strike and dip; (12) sampling sites for fission-track dating; (13) sampling sites for nannoplankton analysis.

Paleogene–lowermost Eocene (Gladenkov *et al.*, 1998).

Recent age determinations for terrigenous rocks of the Karaginskiĭ accretionary prism are based primarily on radiolarians from chert intercalations in sedimentary melange and on nannoplankton from mudstones of the flysch complex. According to Chekhovich *et al.* (1990), cherts in sedimentary melange yield the middle Eocene

radiolarians. The middle–late Eocene and likely associated early Oligocene forms are identified in nannoplankton from the flysch complex (Shcherbinina, 1997). Analogs of the flysch complex of the Govena Peninsula are attributed to the Eocene–Oligocene (Chekhovich *et al.*, 1990; Chamov, 1996). Hence, the sedimentary melange may be slightly older than the flysch complex. The dating results are based, however, on few different plankton assemblages and should be verified therefore.

FISSION-TRACK DATING OF DETRITAL ZIRCONS

The zircon fission-track dating (Wagner and Haute, 1992) is suitable for age determination of barren sequences and stratigraphic correlation. It also can be used for identification of terrigenous material provenances and in analysis of rocks exhumation processes (Garver *et al.*, 1999; Garver *et al.*, 2000a; Garver *et al.*, 2000b). Permitting age determination of separate mineral grains, the fission-track dating offers a possibility to separate zircon populations dissimilar in age and related to different provenances. The cooling of rocks in provenances is under control of different geological processes. On the one hand, volcanic rocks and near-surface intrusions get cool and eroded quickly, and their zircons quickly appear in the sedimentation basin thus being applicable for dating of barren terrigenous sequences (Brandon and Vance, 1992; Garver and Brandon, 1994; Garver *et al.*, 2000b; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). On the other hand, the rock blocks rising up from deep levels are cooling (Wagner and Haute, 1992; Garver *et al.*, 1999) and cross at a certain moment the temperature level of the track system closure in zircons (215°–240°C, Brandon and Vance, 1992). Tracks formed and accumulated in zircon crystals since that moment can be used therefore to determine the age of rock cooling.

Ages of detrital zircons from terrigenous deposits usually vary greatly. For our purpose, it is necessary to establish the youngest population or the minimal age (Garver *et al.*, 1999) determining the lower time limit of sedimentation, as deposits are always younger than their constituent clasts. We used this approach for dating terrigenous deposits of the Karaginskiĭ Island.

Rock samples (6–8 kg in weight) were collected from the Supra-Khynkhlonai sequence (Sample Ka1, psammitic tephroid) and from flysch (samples Ka2, Ka7, Ka11, Ka13, Ka15, sandstones; Ka9, fine-pebbly conglomerates), matrix (samples Ka17, Ka22, Ka28, sandstones), and blocks (samples Ka18, Ka20) of sedimentary melange complex (Fig. 1, Table 1). Zircon grains were separated based on the standard technique in the Laboratory of Accessory Minerals at the Institute of Lithosphere of Marginal Seas, Russian Academy of Sciences (IL RAS). The zircon fission-track dating was carried out at the Union College (Schenectady, N. Y., USA) and the IL RAS. Some peculiarities of the method are described in the footnote to Table 1. From five to fifty zircon grains were dated per each sample. The analyzed samples yield widely ranging ages of individual zircon grains (Fig. 2). This suggests that after deposition the rocks have not been heated above the temperature (215°–240°C) of the track system closure in zircon (Brandon and Vance, 1992). The same is evident from lithology and structure of studied terrigenous deposits. They are lacking secondary minerals formed at the temperature exceeding 200°C and the cleavage formed under certain P/T conditions. Hence, fission-track dating of zircon grains from the sequences studied reflects their cooling times in provenance, and the sedimentary sequence is not older than the zircon grains it bears. The fission-track age of the youngest zircon population defines the lower limit of deposition time of host sandstones.

The age of the youngest population of zircon grains is timing sedimentation, if synchronous volcanism took place in the immediate vicinity during the sedimentation (Brandon and Vance, 1992; Garver and Brandon, 1994; Garver *et al.*, 2000; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). In orogenic belts experiencing a speedy rise and erosion, the near-surface intrusions are quickly subjected to washout at the surface. Hence, the time span between zircon crystallization in intrusion and its appearance in sediments of the adjacent basin is not greater than few m.y. (Shapiro *et al.*, 2001).

One sample of psammitic tephroid (Ka1) was taken from an interlayer about 40 cm thick in the Supra-Khynkhlonai sequence. The rock is composed of dominant volcanic clasts, plagioclase, and insignificant admixture of pyroxene and ore minerals. Volcanic fragments are similar in composition but vary in structure from vitrophyric to lathy; light-colored microlitic rocks prevailing. Most zircon grains belong to population 50 Ma old, and their age corresponds to the boundary between the Ypresian and Lutetian (Table 1).

Five sandstone samples (Ka2, Ka7, Ka11, Ka13, Ka15) and one sample of fine-pebbly conglomerate (Ka9) were taken from the flysch complex. The composition of sandstones is common for flysch: feldspar, quartz, acid volcanics, and mudstones dominate clastic material. All the samples yielded two to three populations of detrital zircons, the youngest of which range in age from 45 to 30 Ma. The indicated time span corresponds to the second half of the Lutetian, Bartonian, late Eocene, and early Oligocene. The young zircon population (40 Ma) from fine-pebbly conglomerate, which is composed mainly of acid volcanics, is of the Bartonian age (Table 1).

Ages of zircons from rocks of sedimentary melange complex vary within a wide range. Sandstones similar in composition to sandstones of the flysch complex are composed predominantly of plagioclase and quartz (Ka17 and Ka28). Young zircons are not abundant in these rocks (10–20%), and their populations ranging in age between 18 and 24 Ma correspond to the early

Table 1. Fission-track ages of detrital zircon from terrigenous and tuffaceous rocks of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka)

Sample no.	Group, formation	K	Age of zircon population		
			P1 (Ma)	P2 (Ma)	P3 (Ma)
Ka28	Melange	30	18.6 ± 3.5 (19%)	39.9 ± 3.1 (60.8)	88.0 ± 9.9 (20%)
Ka17	Melange	20	23.7 ± 3.9 (10%)	57.3 ± 5.2 (62%)	116.0 ± 23.9 (28%)
Ka22	Melange	40	26.1 ± 1.5 (57%)	56.0 ± 10.7 (18%)	96.4 ± 14.2 (25%)
Ka13	Flysch	35	29.9 ± 3.6 (47%)	54.7 ± 5.2 (53%)	
Ka15	Flysch	31	30.4 ± 1.6 (87%)		117.6 ± 14.8 (13%)
Ka11	Flysch	40	36.1 ± 2.4 (40%)	66.4 ± 5.9 (24%)	111.0 ± 8.4 (36%)
Ka9	Flysch	50	39.7 ± 2.6 (56%)	62.3 ± 8.9 (35%)	94.6 ± 39.0 (9%)
Ka7	Flysch	35	44.2 ± 2.6 (68%)	95.5 ± 7.3 (32%)	–
Ka2	Flysch	12	45.6 ± 4.6 (66%)	93.4 ± 14.5 (34%)	–
Ka1	Supra-Khynkhlonai sequence	30	50.2 ± 3.2 (93%)	–	188.3 ± 52.7 (7%)
Ka20	Block in melange (?)	35	70.1 ± 4.4 (80%)	97.5 ± 13.3 (20%)	–
Ka18	Block in melange (?)	5	82.1 ± 10.4 (100%)		

Note: K, amounts of dated zircon grains in sample. P1, P2, P3, zircon populations established on the bases of the program BinomFit v 1.8 (Brandon, 1992; Brandon, 1996). Ages are given in Ma, error corresponds to ±1σ, percentage in parentheses are calculated of the given zircon population relative to the total number of grains dated (K). Zircon grains were dated by the external detector method (Wagner and Haute, 1992), the procedure is described by Garver *et al.* (2000). Zircon grains were pressed into FEP Teflon^{MT} plates 2×2 cm^2 in size. Two plates were prepared for each sample. The plates were roughed out with emery cloth (800 grit) and then polished with diamond paste (9 μm and 1 μm), and Al_2O_3 paste (0.3 μm) at the final stage. Chemical etching was carried out in NaOH–KOH solution at the temperature of 228°C during 20 (the first plate) and 26 hours (the second plate). After etching, the plates covered with detector (mica with a low U content) were exposed to the thermal neutron flux of about 2×10^{15} neutron/cm^2 (thermal reactor of the University of Oregon) simultaneously with the zircon standard samples FCT (Fish Canyon Tuff) and BL (Buluk Tuff), and with glass-dosimeter with known U content (CN-5) (Hurford, 1998). The fission tracks are calculated under microscope Olympus BH-P with automated system and digital plotter (dry method, maximal magnification 1600). The ζ-factor (Hurford, 1998) calculated for age standards (FCT, BL) was found to be equal to 346.22 ± 9.57 (A.V. Solov'ev) and 355.03 ± 8.16 (J. Lederer).

Miocene. Gravelstone (Ka20) of a similar composition consists of quartz, feldspars, their aggregates, and of abundant chert clasts. The youngest population of detrital zircon from the rock is 70 Ma old (Maastrichtian). A peculiar tuffaceous sandstone (Ka22), the main components of which are large angular plagioclase grains and clasts of deformed mudstone, yielded the young zircon population 26 Ma old (late Oligocene). Lastly, zircons from lithoclast tuff (Ka18) with pyroxene admixture represent one population (Table 1) that is 82 Ma old (the earliest Campanian).

NANNOPLANKTON

We collected mudstone samples from the Supra-Khynkhlonai sequence, the flysch and sedimentary melange complexes for nannoplankton study and geochemical analysis aimed to establish the source of clay material. In total, four sample sets have been collected (Fig. 1, Tables 2, 3).

The first collection includes 29 samples (A–Z) of mudstones conformably overlying the Khynkhlonai Formation in upper courses of the Severnaya River. Mudstones enclose here thin tuff and tuffite interlayers. The tuffite Sample Ka1 was taken from this exposure for zircon dating. The ideally exposed section sampled from top to base is 120 m thick.

Sample F, the most representative, yielded one specimen of the late Campanian–early Maastrichtian *Reinchardtites levis* associated with smaller *Reticulofenestra* spp., *R. dictyoda, Chiasmolithus consuetus, Lophodolithus nascens*, and *Coccolithus pelagicus* characterizing altogether the late Ypresian–early Lutetian interval (52–45 Ma; Fig. 3). Due to predominance of Eocene species, the Late Cretaceous form is regarded as redeposited. Another six samples bear single nannoplankton forms. These are *Reticulofenestra* cf. *minuta* (Bartonian–early Oligocene) from Sample C, *Chiasmolithus* sp. (Paleocene–early Oligocene) from Sample J, *Chiasmolithus solitus* and *Reticulofenestra dictyoda* (late Ypresian–Bartonian) from Sample K, and *Coccolithus pelagicus* (Paleocene–present-day) from Sample S. Sample I yielded one specimen of *Micula staurophora* characteristic of the Coniacian–Maastrichtian interval. This form is likely redeposited. The Paleogene age of species from this series of samples is consistent with the early Ypresian–early Lutetian age of nannoplankton from Sample F.

The second collection (H1–12, 005–1/3, 4, 5) represents mudstones of the flysch complex. It characterizes

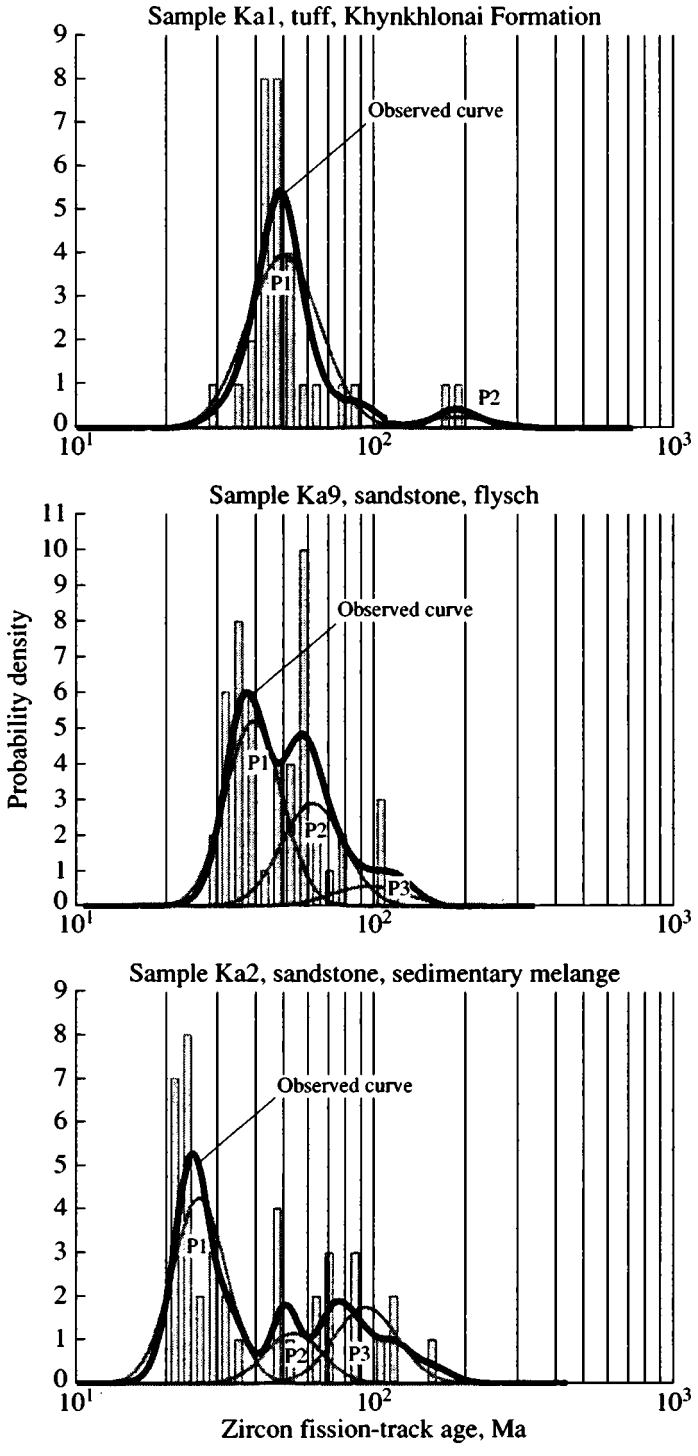

Fig. 2. Fission-track dating results for detrital zircons from tuffaceous–terrigenous deposits of the Karaginskiĭ Island (eastern Kamchatka); P1, P2, P3, peaks of different-age zircon populations (see Table 1) established using program BinomFit 1.8 (Brandon, 1996).

a profile across the flysch complex. Nannofossils found in seven of fifteen samples are exclusively the Mesozoic forms: Bajocian–Maastrichtian *Watznaueria barnesae* (005–1/5), Coniacian–Maastrichtian *Micula staurophora* (H5, 12), Campanian *Broinsonia parca constricta* and *Ahmuelerella regularis* (H8), Turonian–Campanian *Reinchardtites anthophorus* (H9), and Turonian–Santonian *Micula staurophora, Eprolithus floralis, Eiffellithus turriseiffeli, E. eximius, Watznaueria barnesae*, and *Retecapsa ficula*, (H10). Although

Table 2. *In situ* nannoplankton from terrigenous deposits of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka); data of Shcherbinina (1997) included (see Fig. 3)

In situ forms	Khynkhlonai Formation	Flysch complex	Sedimentary melange complex
Lophodolithus nascens	+	–	–
Chiasmolithus consuetus	+	–	–
C. solitus (CP13, 14, Kronotskii Pen.)	+	–	–
Chiasmolithus sp.	+	–	–
Reticulofenestra dictyoda	+	–	–
Reticulofenestra cf. *minuta*	+	–	–
R. umbilicus (CP14, Baklanov Fm.)	–	+	+
R. hillae (CP14, Kronotskii Pen.)	–	+	–
R. oamaruensis	–	+/–	–
R. coenura (CP13, 14, Kronotskii Pen., Baklanov Fm.)	–	+	–
Dictyococcites bisectus (C14, Kronotskii Pen.; CP14–16, Govena Pen., Baklanov Fm.)	+/–	+	+
D. scrippsae (CP14, NW Govena Pen., Baklanov Fm.)	–	+	–
D. daviesii (CP14, Kronotskii Pen.)	–	+	–
Cyclicargolithus floridanus (CP13, 14, Kronotskii Pen., Baklanov Fm.)	–	+	+
C. abisectus (CP14, NW Govena Pen.)	–	+?	–
Discoaster deflandrei (CP14–16, NW Govena Pen., Baklanov Fm.)	–	+	–
D. saipanensis (index species of the CP14b Zone)	–	+	–
Isthmolithus recurvus (index species of the CP 15b Zone)	–	+	–
Sphenolithus moriformis (CP 11, 13, Kronotskii Pen.; CP 14, NW Govena Pen.)	–	+	+
S. predistentus	–	–	+
Coccolithus pelagicus			
C. floridanus	–	+	–
Cribrocentrum reticulatum	–	+	–

Table 3. Redeposited nannoplankton from terrigenous deposits of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka), data of Shcherbinina (1997) included

Redeposited forms	Khynkhlonai Formation	Flysch complex	Sedimentary melange complex
Reinhardtites anthophorus	–	+	–
R. levis	+	–	–
Micula staurophora	+	+	+
Watznaueria barnesae	–	+	+
Broinsonia parca constricta	–	+	+
Ahmuelerella regularis	–	+	–
Eprolithus floralis	–	+	+
Eiffellithus turriseiffeli	–	+	+
E. eximius	–	+	+
Retecapsa ficula	–	+	–
Arkhangelskiella cymbiformis	–	–	+

our samples from the flysch complex yielded only Cretaceous forms, previously the early Eocene nannoplankton was also found in the same sequence (Shcherbinina, 1997). Moreover, the fission-track ages of youngest non-overheated zircon grains from sandstone and gravelstone of that sequence correspond to the mid-

dle Eocene–early Oligocene (46–30 Ma). The zircon dating discredits the pre-middle Eocene age of the sequence, and we believe the Cretaceous nannofossils from flysch samples are most likely redeposited.

The third collection (H13–33) characterizing aleuropelitic matrix of sedimentary melange complex includes only three samples with nannofossils (16, 27, and 33). They yielded the Coniacian–Maastrichtian *Watznaueria barnesae, Micula staurophora*, and *Eiffellithus turriseiffeli.*

The fourth collection (Ha21/ A–L) characterizing matrix of a pebbly mudstone lens in the sedimentary melange complex is sampled from different parts of a small (20 × 30 m) exposure on the watershed. Three of eleven samples yielded only Mesozoic nannofossils: Campanian *Micula staurophora, Arkhangelskiella cymbiformis, Broinsonia parca constricta* (I), Bajocian–Maastrichtian *Watznaueria barnesae* (K), and Cretaceous forms undeterminable at the species level (Z). One sample (V) yielded the Paleogene (Eocene–Oligocene) forms *Cyclicargolithus floridanus* and *Sphenolithus moriformis* (Fig. 3). The most interesting is Sample A, in which Cretaceous *Eiffellithus eximus* and *Eprolithus floralis* (Turonian–Santonian) are joined together in a pellet, whereas the clayey matrix bears *Reticulofenestra umbilicus, Coccolithus pelagicus, Dictyococcites bisectus*, and *Sphenolithus predistentus*, the assemblage of which characterizes the Bartonian–earliest Oligocene interval (41–28 Ma) (Fig. 3). It is apparent that this assemblage determines age of pebbly mudstones and thus of a substantial part of the sedimentary melange complex.

DISCUSSION

Redeposition of nannoplankton. The established two nannoplankton assemblages, one redeposited Late Cretaceous and the other one from the Eocene–Oligocene interval of the studied section represent the unexpected result of our study. In abundance and preservation state, the redeposited older forms are comparable with younger ones, and their occurrence frequency is even higher. For instance, exclusively Cretaceous forms are established in mudstone samples from flysch. Only the zircon fission-track dating of flysch sandstones and earlier found Eocene–Oligocene nannoplankton (Shcherbinina, 1997) make us consider the Late Cretaceous forms as redeposited.

In addition to dominating Eocene forms, Late Cretaceous nannofossils were also found in mudstones of the Supra-Khynkhlonai sequence. In one sample (F), Late Cretaceous species occur in association with Cenozoic forms. In Sample Ha21A, from matrix of pebbly mudstones in the sedimentary melange, the Turonian–Santonian species were also found along with Bartonian–early Oligocene forms. In the last case, Cretaceous forms were found inside a pellet of the biogenic origin.

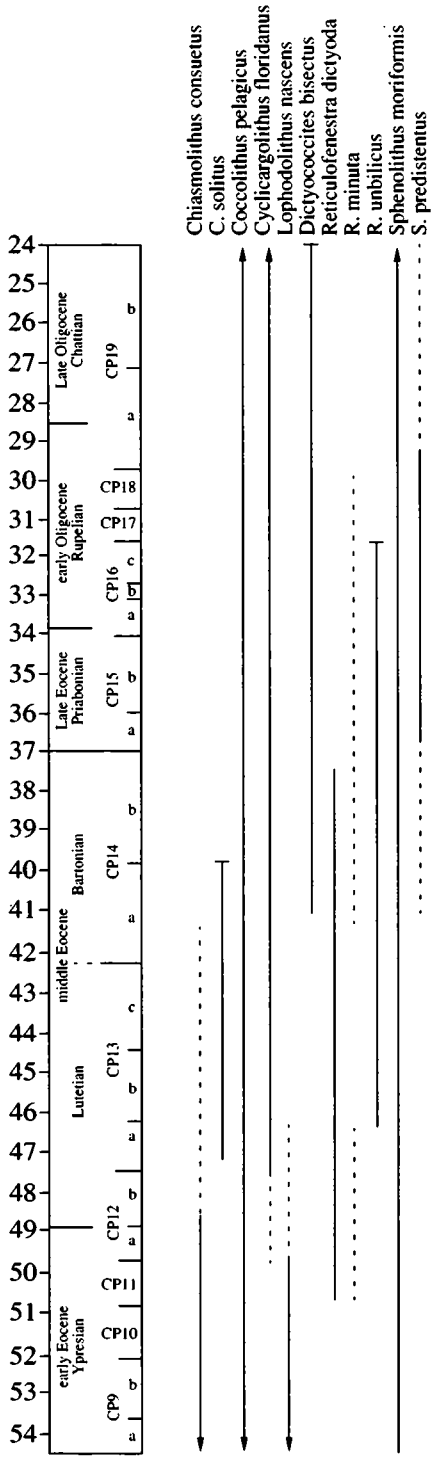

Fig. 3. Stratigraphic range of nannoplankton species found in terrigenous deposits in the northeastern part of the Karaginskiĭ Island (eastern Kamchatka).

The early–middle Eocene nannoplankton assemblage from the Supra-Khynkhlonai sequence substantially differs from the Late Eocene–early Oligocene assemblage found previously in the flysch complex

(Shcherbinina, 1997). The *in situ* nannoplankton from sedimentary melange, which was found in two samples from one and the same exposure, does not differ from nannoplankton of the flysch complex and corresponds to the Late Eocene–early Oligocene stratigraphic range.

The source of redeposited nannofossils and modes of their redeposition represent still an open issue. It is apparent that they are from some Late Cretaceous, mainly Santonian–Campanian and most likely poorly lithified sequence, in which a considerable part of nannoplankton was buried in pellets transportable after washing-out. The Vatyna and Irunei siliceous–volcanogenic sequences of Kamchatka and southern Koryak Upland, which are exposed in the vicinity of Karaginskiĭ Island, are mainly of the Santonian–Campanian age. After the arc–continent collision in the second half of the Eocene and Oligocene, the sequences were definitely subjected to intense erosion. However, nannoplankton has not been found in these strata. The provenance of sandstones and pebbles in flysch and melange, or origin of basaltic olistoliths in flysch and of tuff-basaltic lenses in melange, which are widespread in the southeastern part of the Karaginskiĭ Island, are unknown so far. It is likely that the Late Cretaceous redeposited nannoplankton occurring in the Eocene–Oligocene mudstones of the island was also related to a provenance undeterminable at present.

Hence, the case of Eocene–Oligocene deposits of Karaginskiĭ Island shows that redeposition of nannoplankton from older deposits to considerably younger ones is not only possible theoretically, but can be a significant event. If the nannoplankton content in rocks is low and number of collected samples is small, such a redeposition may cause large errors in dating the terrigenous sequences.

The fission-track dating of detrital zircon. The study carried out confirms in general the previous inference that ages of youngest zircon populations characterize, as a rule, stratigraphic ranges of their host sandstones, which accumulated in tectonically active zones (Brandon and Vance, 1992; Garver and Brandon, 1994; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). In particular, the zircon ages of tephroids from the Supra-Khynkhlonai sequence (50.2 ± 3.2 Ma) are concordant with the sequence age established based on nannoplankton (the late Ypresian–early Lutetian). The age of tuffaceous sandstone from sedimentary melange (26.1 ± 1.5 Ma, Sample Ka22) is very close to the age of nannoplankton from matrix of adjacent pebbly mudstones (Sample Ha21, Bartonian–early Oligocene). The age of most zircon grains in sandstones from the flysch complex falls within the interval of 46–30 Ma and corresponds to the flysch age based on nannoplankton (Shcherbinina, 1997). On the other hand, the young zircon population in the melange complex is younger (26–18 Ma) than the complex age established on the basis of radiolarians (Chekhovich *et al.*, 1990). This is likely related to the fact that cherts in the melange complex are represented by olistoliths, which are older than the matrix.

The age similarity between young zircon populations and sandstones enclosing them implies that the zircon grains are genetically related with either the synchronous volcanism or the very fast exhumation and cooling of intrusions in the provenance. Some sandstones from sedimentary melange (e.g., Ka22) do contain substantial tuffaceous admixture, whereas most sandstones from the flysch complex of the Karaginskiĭ accretionary prism lack any tuffaceous material. More than half of them is made up of acid volcanics and granitoids and their derivatives (quartz and plagioclase) (Shapiro, 1984; Shapiro *et al.*, 2000). Hence, acid magmatic rocks of the provenance may be thought to be the sources of zircon grains. The inference validity can be tested, as the zircon fission-track ages are established not only for sandstones but also for gravelstones of the flysch complex, which are composed of acid volcanics, and for rhyolite and granite pebbles and boulders from pebbly mudstone of the flysch and sedimentary melange (Ledneva *et al.*, in press). It has been found that zircon grains in gravelstones are of the same age like those in sandstones. In granite pebbles, zircons are of Paleocene age, whereas rhyolite pebbles from the same rock are of the Maastrichtian age according to zircon dating. The age difference between granites and rhyolites is easily accounted for by necessity of a certain time-interval for granite exhumation. It can be expected therefore that combined washout of these rocks would give two zircon populations: one corresponding to the Cretaceous volcanism and the other one corresponding to the uplift of shallow intrusions at the surface. The integration of both populations into one with the mode of about 65 Ma is also a possibility. A zircon population with the last age approximately is truly established for some sandstones and gravelstones (Ka9, Ka11). Such zircon grains are less abundant than zircons subsynchronous to sedimentation.

Nonetheless, there are lenticular sandstone and gravelstone (Ka20) bodies in the sedimentary melange complex, in which the first zircon population is 70 Ma old. These are either olistoliths of Cretaceous sandstones or fragments of sandy layers similar in age to other sandstones but barren of zircon making up the young population.

The only fission-track age available for tuffs from the sedimentary melange complex (Ka18) corresponds to the beginning of the Campanian (82 Ma). The most probable explanation in this case is the tuff occurrence in the form of large and small olistoliths.

The data obtained confirm the earlier inference (Chekhovich *et al.*, 1990) that terrigenous rocks, which make up the bulk of the flysch and sedimentary melange complexes in the southeastern part of Karaginskiĭ Island, accumulated between the second half of the middle Eocene to the beginning of the Miocene. Sedi-

ments in these complexes become younger from northwest to southeast, i.e. from the continent to the past trench. The youngest fission-track ages were obtained for sandstones from the sedimentary melange complex.

Formation of the accretionary prism. The initial and terminal formation stage of the Karaginskiĭ accretionary prism did not left geological records. Based on the accretionary prism formation model, it may be suggested that imbricated thrust faults and folds in the flysch complex, like the chaotic complexes of sedimentary melange, are subsynchronous to sedimentation. In this case, the upper part of accretionary prism exposed on the island corresponds to the middle Eocene–early Oligocene subduction under the Kamchatka Peninsula. During that time, the Kinkil (or West Kamchatka–Koryak) volcanic belt formed on the isthmus of the Kamchatka Peninsula (Gladenkov *et al.*, 1997; *Explanatory Notes...*, 2000). At the end of the Eocene to the very beginning of the Oligocene, volcanism was likely active in the Govena arc. However, geological connection of the Karaginskiĭ prism with the Govena arc is doubtful. First, it is unlikely that the Govena arc extended to Karaginskiĭ Island since the Oligocene, because this interval in the island section is represented by terrigenous sediments. Second, the Karaginskiĭ prism is situated too close to the axial zone of the island, where the Govena arc extension is assumed to be. The present-day fore-arc structures more than a hundred kilometers wide are situated between the accretionary prism and the volcanic arc.

The youngest sandstones in the sedimentary melange complex enable a suggestion that the Karaginskiĭ accretionary prism was under formation up to the beginning of the Miocene, and developed even later in its underwater part.

CONCLUSION

(1) The studied mudstones of the northeastern part of Karaginskiĭ Island bear two nannoplankton assemblages of a low diversity and abundance. We consider the first assemblage ranging in age from the Ypresian–Lutetian (mudstones overlying volcanics of the Khynkhlonai Formation) to Bartonian–Oligocene (the sedimentary melange complex) as occurring *in situ* and characterizing age of sediments. The second assemblage of Late Cretaceous–Paleocene forms signifies rewashing of older sediments within the unknown provenance. The redeposited nannofossils are comparable with the *in situ* assemblage in abundance, diversity, and preservation state of species.

(2) According to the fission-track dating, the age of youngest detrital zircon population is close to sedimentation time of their host rocks. In the Supra-Khynkhlonai sequence, such a population corresponds in age to the Ypresian–Bartonian boundary that is consistent with the nannoplankton age of Sample F. In the flysch complex, the zircon populations range in age from 46 Ma (Bartonian) to 30 Ma (early Oligocene) that also corresponds to the sequence age based on nannoplankton (Shcherbinina, 1997). Data on the sedimentary melange complex are less compelling, though when it is possible to establish simultaneously the nannoplankton age of mudstones (Ha21, Bartonian–early Oligocene) and the age of young zircon population in adjacent tuffaceous sandstones (Ka22, 26.1 ± 1.5 Ma), the data do not contradict each other.

(3) In the northeastern part of Karaginskiĭ Island, terrigenous deposits become younger in the southeastward direction from the axial zone composed of Cretaceous and early Paleogene island-arc volcanics. Samples taken from the flysch complex are of the Bartonian–early Oligocene age, whereas the fission-track dating of detrital zircons suggests the Oligocene–early Miocene age of sedimentary melange complex. The situation when rocks become younger toward a trench and rock beds dip predominantly in the opposite direction is typical of accretionary prisms.

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Geochronology of Cenozoic Basalts in Mongolia and Their Relationships with Neotectonic Structures

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Abstract—The epiplatform (Upper Cretaceous–Paleogene) and neotectonic (Neogene–Quaternary) evolutionary stages and morphostructures are distinguished in Mongolia. The neotectonic structures are characteristic of the orogenic zone that is free of volcanics (Altay, Hentiy) and of the fore-rift (Hangay, Dariganga) and rift (Hubsugul region) volcanic zones. Seventeen distribution areas of volcanogenic rocks are outlined, and geochronological database that includes about a hundred of K–Ar and ^{40}Ar–^{39}Ar dates obtained in Russia, Austria, Switzerland, the United States, and Hungary is compiled. The established geochronological reference level range from 120 to 30 Ma in the orogenic zone and correspond to 30–31, 28, 15–13, 8–5 in the fore-rift zone and to 28–25, 15–13, 5–3, and less than 1 Ma in the rift zone. The reference isotopic dates obtained for basalts enable the precise (of zone rank) dating of Cenozoic biostratigraphic (mammalian) horizons in Mongolia (Devyatkin *et al.*, 2002). The biostratigraphic horizons and geochronological levels create a basis for remote and interregional correlations.

Key words: basalts, geochronology, stratigraphy, structures, neotectonics.

Mongolia is a country of hundreds extinct volcanoes. Their distribution area extends across the Dariganga Plateau from the southern termination of the Baikal rift zone to northern China. The zone of Cenozoic volcanism is up to 350–400 km wide (between 100 and 115°E). The Dariganga Plateau is situated between 113 and 115°E (Boishenko *et al.*, 1990). Two additional small areas of volcanism are known in the southwest (the Burun-Huray depression) and easternmost part (Halhin Gol River) of the country. Genshaft and Saltykovskiĭ (1990) distinguished 17 individual basaltic fields, which are either spaced, or adjacent to each other (northern and central Mongolia; Fig. 1).

Basalts of Mongolia are well studied in terms of mineral and chemical composition and with respect to nodules of mantle and crustal genesis they bear (Kepezhinskas, 1975; Saltykovskiĭ and Genshaft, 1984, 1985, 1990; Yarmolyuk *et al.*, 1994, 1995: and many other works). These data were used to infer the Earth' crust structure, distribution of magmatic chambers, and dynamics of volcanogenic processes.

Unfortunately, some other aspects of volcanism in Mongolia are known inadequately and should be clarified further. These aspects already mentioned in publications are as follows: (1) geochronology of basalts and (2) relationships between the basaltic volcanism and neotectonics.

The first isotopic date (Mellet, 1968) was obtained for the basalt sample collected in the 1920s from the Shand-Gol Formation of the Lakes' Valley. Afterward, a series of dates was obtained in the period of works (1960s to 1980s) of the Joint Soviet–Mongolian Geological Expedition (Gabuniya *et al.*, 1975; Devyatkin and Smelov, 1979; Devyatkin, 1981; Saltykovskiĭ *et al.*, 1984, 1985; Genshaft *et al.*, 1990; Shuvalov *et al.*, 1985, 1993; Logachev *et al.*, 1982; Ageeva *et al.*, 1988; Amirkhanov *et al.*, 1985). Devyatkin (1981) suggested the general geochronological scheme for basalts of Mongolia and considered their relations with Cenozoic sedimentary horizons. Later on, he correlated geochronological levels of basalts with the paleomagnetic scale and with sedimentary faunal horizons of the Cenozoic (Devyatkin, 1994).

At present, there are known about 100 isotopic dates obtained at different institutions of the former Soviet Union (Institute of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences (IGEM RAS), Geological Institute (GIN) RAS, Institute of the Earth Crust, Siberian Division (IEC SD) SO RAS, GIN Georgia, Institute of Geochemistry SO RAS, Institute of Physics, Dagestan Branch AN SSSR, Institute of Geology AN Tajikistan) and in other countries. They outline the general geochronological succession of volcanic manifestations in Mongolia, some aspect of which need to be revised however.

The best studied in geochronological aspect is basaltic volcanism of the Lakes' Valley, where Austrian researchers (Daxner-Höck *et al.*, 1997; Höck *et al.*, 1999) recently obtained a series of ^{40}Ar–^{39}Ar dates. They established three geochronological reference levels in basaltic sequence. Level A of 30–32 Ma (16 age determinations) and Level B of 26–28 Ma (6 age deter-

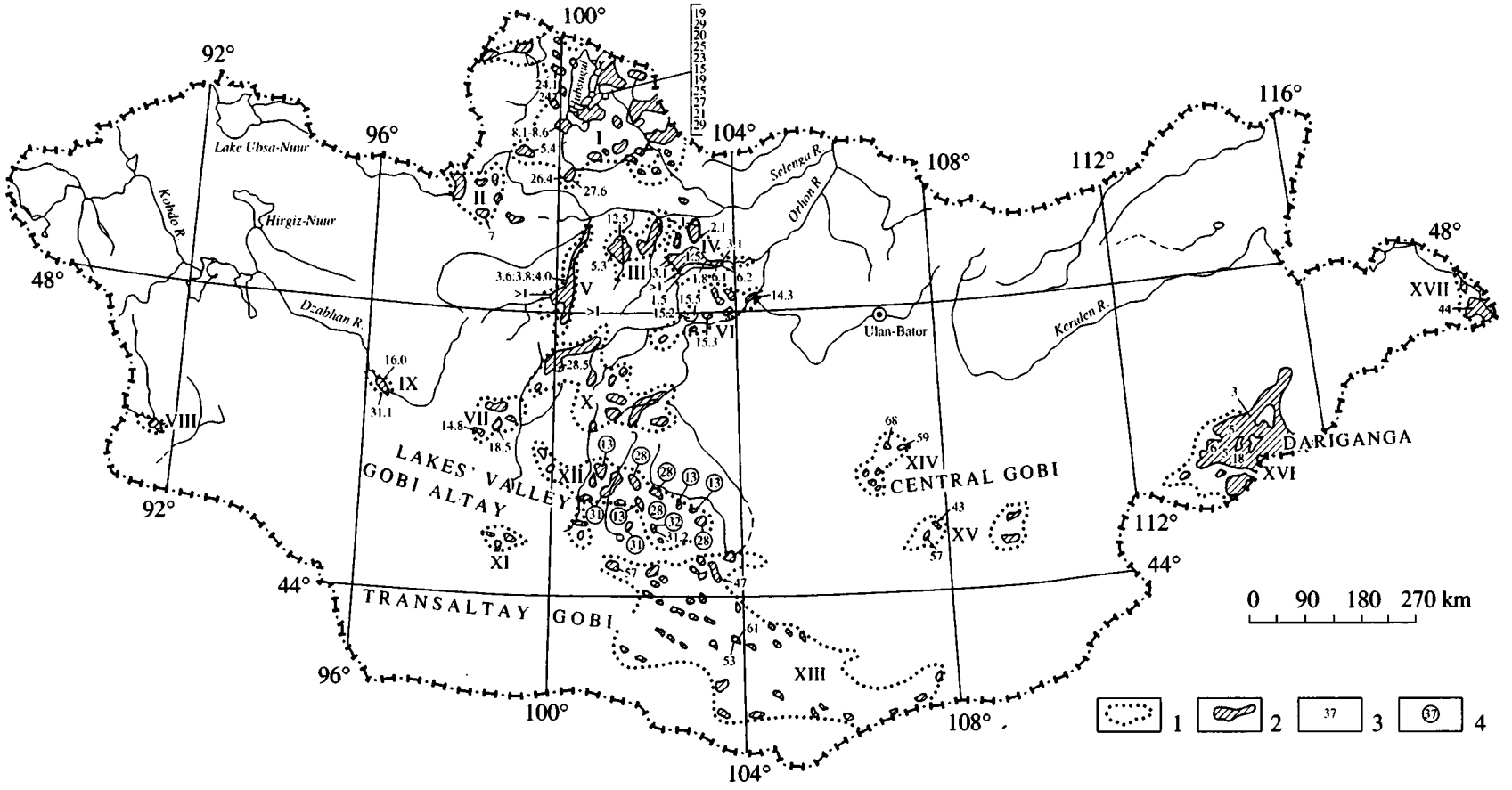

Fig. 1. Areas and isotopic dates of basaltic volcanism in Mongolia (boundaries after Genshaft and Saltykovskiĭ, 1990, with some modifications): (1) basaltic fields; (2) present-day outcrops; (3) K–Ar age, Ma; (4) ^{40}Ar–^{39}Ar age, Ma. Areas by names: (I) Hubsugul; (II) Tesin-Gol; (III) Hanuy; (IV) Orhin–Selenga; (V) Tariat–Chuluut; (VI) Ugey-Nuur; (VII) Baydarig; (VIII) Barun-Huray; (IX) Dzabhan; (X) Hangay; (XI) Gobi Altay; (XII) Lakes' Valley; (XIII) South Gobi; (XIV) Mandal; (XV) Undur-Shili; (XVI) Dariganga; (XVII) Halhin-Gol.

minations) are established at the base of the lower Oligocene Shand-Gol Formation, while Level C of 13 Ma corresponds to the middle Miocene (above the Loo Formation). Using the K–Ar method, we analyzed at the Institute of Nuclear Research of Hungary 7 samples from the same region, and our results are consistent with inferences of Austrian scientists. These and other dates characterizing geochronology of basaltic volcanism in Mongolia are considered below.

Another insufficiently elaborated problem concerns relations between the neotectonics and basaltic volcanism. In available publications, the problem was regarded in a simple generalized manner without due attention to possible links between neotectonic structures of various types and geochronology of volcanic eruptions.

There are several types of structures different in morphology, formation time, and character of volcanism, which appeared during the period of neotectonic activity in Mongolia (*Neotectonics...*, 1998). They are different in age (*Mesozoic...*, 1975; Devyatkin, 1981) and show complex geological relations.

(1) The platform (pre-neotectonic) stage of evolution spans the Cretaceous and Paleogene (the Oligocene should be excluded probably). According to formation criteria, the commencement of neotectonic stage is distinctly confined to the Oligocene–Miocene boundary. However, the areal volcanism of the fore-rift stage was likely triggered already in the Oligocene.

The relief peneplanation at the time was polygenetic, of the denudation type in the north and of the accumulative genesis in the south of Mongolia (*Geomorphology...*, 1978). Basalts of the platform stage are concentrated in southeastern and central Gobi, where they constitute individual massive necks, remains of lava flows, and dike swarms. These areal volcanic manifestations were inherited from the Late Cretaceous volcanic activity and have no connection with neotectonic structures. Manifestations of this kind are known to the east and south of the Gobi Altay and in the Undur-Shili and Mandal areas. Isotopic dates for basalts of the platform stage range between 59 and 43 Ma.

(2) The neotectonic stage is manifested in several zones of different morphology and formation time. The orogenic zone is spread over the Mongol and Gobi Altay, Hentiy, and separating intermontane depressions. The western part of the orogenic belt is amagmatic. In contrast, the southern, plateau and watershed Hangay, the elements of the fore-rift zone, were areas of active volcanism that created different morphostructures, which retain their original morphology even at present. These are remnants of large basaltic fields (hundreds square kilometers), which are composed of several (up to ten) lava flows, linear fissure eruptions, neck bodies, and dike swarms, all confined to activated faults of northwestern (to sublatitudinal) strike. Characteristic of these volcanics are three geochronological reference levels (30–32, 26–28, and 13 Ma), although some dates correspond here to 18, 16, and 14 Ma, thus being between the levels established in the Lakes' Valley. The maximum volcanic activity in the region under consideration was likely in the Miocene.

In the crest area of Hangay (fore-rift zone), there are thick (up to 100 m) sequences of basaltic flows, the base of which is dated back to 28.5 Ma. From the crest area, the flows slope down into valleys to form there the "valley" complex. Flows of the crest area seem to be different in age. On both sides of the Egin-Daba pass (ridge crest), there are two volcanic cones looking like young, Quaternary or even Holocene, which give rise to small lava flows extending for several kilometers down to the valley floor.

In the Hubsugul, Hangay, and Bolnay areas of the fore-rift zone, there are narrow sublatitudinal Tariat and Dashinchilen grabens. Volcanic rocks occupy here separate areas. Geochronological levels of basaltic eruptions established here are as follows. Level A of 14.4–15.5 Ma is established in the northern wall of the Dashinchilen trough. Basalts in the Chuluut, Hanuy, and Orhon river valleys extend from here to the Selenga River and correspond to Level B of 6.3–1.8 Ma. Basalts at the base of this complex are older, dated back to 12.5–11.2 Ma. Level C characterizes basalts of the Tariat rift and of the Orhon, Chuluut and other river valleys. These basalts range in age from about 1 Ma to 6.6 ka near the Lake Terkhin-Tsagaan-Nuur. The northern part of the fore-rift zone (Hubsugul area) is separated from the Hangay by the Selenga River valley free of basaltic volcanism.

In the Hubsugul rift zone, basaltic field is indivisible. Dates of 15–29 Ma characterize basalts of the rift eastern wall and, likely, the basaltic plateau to the east of the Hubsugul Lake. Several lava flows that constitute the plateau sequence are dissected by river valleys being produced by recurrent Oligocene eruptions.

In western and southern Hubsugul areas, remnants of basaltic sheets occur in the crest area of the coastal ridge. The basalts here are approximately concurrent to basaltic lavas of the eastern Hubsugul area (24–27 Ma). It seems reasonable to suggest that before the formation of the Hubsugul rift all the basalts belonged to a single sheet of plateau basalts, which have not been deformed by tectonic movements at that time. At present, the basalts are at the relative altitude greater than 1000 m above the water level of the lake. The commencement of rifting was likely in the terminal Miocene (8 to 5 Ma), when lavas flew along the valleys east of the lake. Now they are risen up at the absolute altitude of 2500–2700 m. The Hubsugul and Darhat rifts proper are free of recent basalts.

On the Dariganga Plateau corresponding to the other portion of the fore-rift zone, basaltic volcanism commenced in the middle Miocene, as it is evident from stratigraphic data (Devyatkin, 1981) and from an isotopic date of 18.0 Ma characterizing the base of the plateau basalts (Genshaft *et al.*, 1990). The second

Miocene–Pliocene complex (3 to 6 Ma) includes upper flows of plateau basalts and some volcanoes situated above them. The third Quaternary (locally Holocene) complex corresponds to well-preserved scoriaceous lavas. The Dariganga Plateau escaped the rifting stage despite a considerable tectonic faulting.

I. Hubsugul area. Basalts are exposed on both sides of the Hubsugul Lake. Large and thick (100 to 200 m) basaltic plateaus on the east are dissected by river valleys. Being faulted along the lake coast, basalts are concealed under water. On the west, small, though thick (tens to hundreds meters) basaltic remnants are exposed in the crest of the Hubsugul Ridge (altitude from 2700 to 2800 m). This complex is 24.1–24.4 (±0.5) Ma old. Basalts, which are located to the south of the Hubsugul trough in the Egin-Gol–Muren interfluve and on flanks of the ridge (altitude 1940 to 2000 m), likely belong to the same complex, being 27.4–27.8 (±0.6) Ma old.

The second complex corresponds to small basaltic remnants located to the southwest of the Hubsugul Ridge at the altitude of 2550–2650 m. These lava are 8.1–8.3 (±0.2) Ma old.

The third youngest complex is established in the Arasain-Gol, Tsagaan-Gol, and Han-Gol river valleys southwestward of the Hubsugul Ridge. This is the valley generation of high terraces (altitude of 1980–2260 m). The basalts are 5.4 ± 0.5 Ma old.

The first complex was formed prior to the rifting stage. At that time, before the rift graben formation, a single volcanic sheet united basalts of eastern and western rift walls. Younger basaltic complexes erupted at the rifting stage, when activated meridional (axial) faults prograded southward and caused uplift of the West Hubsugul Ridge for 1500–2000 m at least (the summary amplitude during the entire neotectonic stage).

Thus, there are three basaltic complexes in the Hubsugul rift zone:

(1) The pre-rift watershed complex includes basalts of western and eastern flanks of the present-day rift, which formerly belonged to one lava field. Two or more phases of volcanic activity are likely recorded in this lava sequence.

(2) Basalts as old as 8.8–8.1 Ma occur in isolated areas confined to activated faults of the initial rifting stage. Their distribution is limited by a meridional fault system extending to the south from the Hubsugul graben. The basalts are characteristic of low (2300–2500 m) watersheds. The complex definitely belongs to the rift stage that activated meridional faults and caused the graben formation.

(3) Basalts of the rifting stage proper (5.4 Ma) are of a limited distribution in the western periphery of the Hubsugul rift.

Unfortunately, basalts of the Hubsugul rift eastern wall are poorly studied. They are as thick as 50 m along the lake coast and form thicker (100–200 m) plateaus eastward, which are dissected by river valleys. As is seen in coastal cliffs, the complex includes 2–3 lava flows locally separated by weathering crust. Southward of the Hubsugul Lake, the flows overlie red beds (clay and sand) resting on the peneplanation surface of basement rocks. K–Ar dates, which have been obtained at the Institute of Physics, Dagestan Branch AN SSSR, for 12 basalt samples form coastal exposures, range between 19 and 29 Ma (average value 24 ± 5 Ma). Asynchronous eruptions within this time span cannot be excluded however.

In my opinion, western and eastern basalts belonged to a single lava field before the rifting stage. The meridional fault system coinciding with the Hubsugul Lake axis was activated in the early Miocene. This event was responsible for the axial graben formation, disruption of lava field, and for the western flank uplift. The process continued throughout the Pliocene and is likely in progress even at present. During the neotectonic stage as a whole, the axial part of western flank was risen up to 1500 m above the lake surface and up to 1800 m above the lake bottom.

II. Tesin-Gol area have not been studied in the geochronological aspect. On the left side of the Tesin-Gol River, plateau basalts are as thick as 200 to 300 m. In northerly localities, they are overlain by fresh looking lava flow and crowned by two young (Holocene) cinder cones up to 100 m high. Small basaltic plateaus are known in the east of the area.

III. Hanuy area is situated near the Hajrhan and Bayan Agt in middle courses of the Hanuy River. It is likely an individualized eruption center of flat lava flows adjusted well to the former erosional relief of medium-high mountains, which seems to be relatively young therefore. The lava flows that flooded the relief have a smooth surface, being crowned by several cinder and lava cones (up to 100 m high) with well-preserved craters. In the Hanuy River valley, the lava flow extends down to its junction with the Selenga River, being overlain by pebbly terrace deposits. Characteristic of lava flows is columnar jointing. A small lava flow of the Tuldura-Gol River likely belongs to the same area. In the riverhead, there are two young volcanic cones corresponding in age to 12.5 ± 1 and 5.3–5.4 (±0.3) Ma.

IV. Orhon–Selenga area extends from middle courses of the Orhon River into valleys of its left-side tributaries. Basaltic lavas fill in the erosional valleys and depressions. The basalts are well preserved. Several lava cones up to 150 m high are also known in the area. Descending via tributary valleys, lava flows reach the Orhon River valley, in which they are traceable for a distance of 35–40 km.

Several lava horizons different in age occupy concrete geological and geomorphological positions in the Orhon–Selenga area. The oldest complex corresponds to fragmented terraces on the left side of the Orhon River (altitude of 1160–1220 m), which are frequently overlain by younger basaltic flows. According to K–Ar dates, the complex is 11.5 ± 0.2 Ma old.

The next complex up to 100 m thick is established at the right side of the Orhon River, where basalts from the base of the 120-m-high terrace are 4.0 ± 0.5 Ma old (Devyatkin, 1981). Samples from the top of the same terrace yielded the K–Ar date of 3.1 ± 5 Ma (Gabuniya *et al.*, 1975; Devyatkin, 1981). In addition, wall rocks of the preserved volcano Tsagaan-Dzurihe, which have been analyzed at the Geological Institute RAS, yielded the date of 2.1 ± 0.1 Ma.

A series of K–Ar dates characterizes basalts exposed lower in the Orhon River valley (Devyatkin, 1981), where two lava flows are recognized. The older flow at the top of 90-m-high terrace is overlain and underlain by red beds bearing bones of the middle Pliocene mammals (Devyatkin, 1981; *Late Cenozoic...*, 1989). This flow is 3.1 ± 0.5 Ma old according to the first age determination (Gabuniya *et al.*, 1975), but subsequent measurements yielded the older values of 6.2 to 6.4 ± 0.5 Ma (Devyatkin and Smelov, 1979).

The second flow forming a low inset terrace (20–30 m high) is 1.8 ± 0.1 Ma old (Devyatkin, 1981). The large Tsagaan-Dzurihe Volcano situated at the right side of the Orhon River was likely one of eruption centers of the Orhon–Selenga area. It is a large neck body of a complex structure up to 10 km in diameter and rising up for about 1500 m above the riverbed (altitude of 2077 m). K-Ar dating of rocks from the volcano top and 10-m-high river terrace showed that volcano is younger than 1 Ma.

Thus, there are four chronological levels of basaltic eruptions in the area under consideration. They correspond to 11–12 Ma (middle Miocene), 3–6 Ma (middle Pliocene), 1.8–2.0 Ma (late Pliocene), and to less than 1 Ma (Pleistocene). This succession of volcanic pulses is similar to that of the Hanuy area.

V. Tariat–Chuluut area is situated in the Central part of the Hangay Ridge and corresponds to the eastern part of a latitudinal graben. Lava flows of the Chuluut River valley almost reached the Selenga River. K–Ar and other dating methods (Devyatkin, 1981; Logachev *et al.*, 1982) revealed the following three complexes of basaltic lavas:

(1) The Pliocene complex of high terraces in the eastern part of the Terhin–Tsagaan-Nuur depression is 200 to 225 m thick and includes 10 lava flows intruded by dikes (Kepezhinskas *et al.*, 1975; Kepezhinskas, 1979). Basal horizons of the complex are composed of tuff-breccias and scoriaceous lavas. Downstream of the Sumein-Gol River joining the Chuluut River, the complex corresponds to remnants of a large volcanic edifice representing an oval up to 0.7 km in diameter. The edifice rising up for 100 m above the terrace is composed of scoriaceous lavas. Steep cliffs of the high terrace on the right bank show that scoriaceous lavas and crosscutting dikes are overlain by basaltic flows with columnar jointing. This volcanic center functioned several times.

Downstream, along the Chuluut River, terraces decrease in height to 100–120 m and depict the geomorphological surface of several lava flows, and then, 20 to 30 km away, the terraces are crowned with denuded volcanic centers raising as 50-m-high stocks above the surface. As is seen in terrace outcrops, the stocks composed of lavas with columnar jointing are as thick as 50 m and more.

Below and inside the Pliocene lavas, there are lenses (3 to 5 m thick) of ferruginate pebblestones and gravelstones. Except for one lava flow in the lower part of the complex, the other rocks reveal the reversed magnetic polarity.

The isotopic dating of basalts that has been carried out at the Institute of Geochemistry in Irkutsk yielded the dates of 4.0 ± 2.5, 3.8 ± 0.2, and 3.6 ± 0.2 Ma, which characterize, from the base upward, the succession of basalts in one terrace (120–140 m high) of the Chuluut stream meander (Logachev *et al.*, 1982). The dates suggest the early Pliocene age of the basalts (Matuyama Chron).

(2) The Quaternary complex of medium-high (70–100 m) terraces is widespread in the eastern part of the Terhin-Tsagaan-Nuur depression and extends from here into the Chuluut River valley. Volcanoes confined to the fault system around this depression represent the eruption centers of Quaternary lavas. They are eroded to a different extent, thus being different in age. The Shavarin-Zaram Volcano is morphologically indistinct. The watershed, where the volcano is situated, is overlain by massive and scoriaceous lavas bearing nodules of crustal and mantle rocks. The complex has not been dated properly. According to unchecked information, the Shavarin-Zaram Volcano is 1.0–1.2 Ma old.

Sections of the medium-high terraces (70–100 m) of the Chuluut River show that basaltic lavas are intercalated with interlayers and lenses of pebbly to gravely deposits. All lava flows of the intermediate complex have normal magnetic polarity and correspond to the Brunhes Chronozone. The TL date of 520 ± 25 ka has been obtained at the Geological Institute of Ukraine for basal sediment burned at the contact with lavas. The other TL dates characterize the overlying (130 ± 17 and 180 ± 20 ka) and inset (200 ± 25 and 280 ± 30 ka) lacustrine deposits of the basaltic terrace situated upstream of the bridge across the Chuluut River (Logachev *et al.*, 1982). These dates imply the Quaternary age of the complex that is of a multiphase character, as one can judge from a variable preservation state of volcanic edifices and from lava interlayering in terrace sections.

(3) The Holocene complex of the Tariat Volcano is distinguished in the western area part. The volcano representing a cinder cone with basal diameter of 500 m is 200 m high and well preserved. The southern side of its crater is cut through by lava channels, and relevant lavas dammed the Lake Terhin-Tsagaan-Nuur. On the eastern side, the Holocene lava flow of the Sumein Canyon extends for a distance of 40 km toward the point of junction with the Chuluut River. This flow retained flu-

idal patterns at the surface. The eastern base of the volcano is composed of black ash, scoriaceous lavas, and volcanic bombs up to 1 m across. Lava bubbles near the lake are partially destroyed. It seems that lavas were hot, when they dammed the lake. There is a legend that the Tariat Volcano was smoking in historical times, and this seems probable, because the edifice is well preserved.

The Holocene lavas have not been dated, and their age is inferable by implication from sediments sampled at the lake bottom near the volcanic dam. The radiocarbon dating of organic matter in sediments from the depth of 6.1 m showed that the matter is 6890 ± 400 years old (Doroveiuk and Tarasov, 1990). In the Tariat depression, there are two Holocene volcanoes: one near the southern wall and the other one near the bridge across the Chuluut River. They erupted flows of very fresh lavas with columnar jointing, which are 5 to 7 m thick only.

VI. Ugey-Nuur area extends from the Tsetseleg Mountains in the northeast to the junction of the Orhon and Tola rivers. Several volcanic centers of the area (large neck bodies) erupted lava flows, which slope down to the south toward the Dashinchilen latitudinal trough and frame its northern wall. In opinion of Saltykovskiĭ *et al.* (1984), these were fissure eruptions. The eruption sites (Ihe-Togo-Bogdo, Bagen-Darhu-Ula, and others) are marked by accumulations of scoriaceous lavas, pyroclasts, and crustal xenoliths. Four K–Ar dates characterizing the eastern part of the area are as follows: 14 ± 4 Ma (data of I.K. Volchanskaya), 15.5 ± 2 Ma (two whole-rock samples), and 15 ± 2 (sanidine fraction; Saltykovskiĭ *et al.*, 1984).

V.F. Shuvalov and T.V. Nikolaeva from the IGEM RAS sampled basalts of the lower plateau in the Tola–Harduhin Gol interfluve and determined their age that is equal to 14 ± 3 Ma. Thus, volcanism of the Ugey-Nuur area took place in the narrow time span (14–15 Ma) of the middle Miocene.

VII. Baydarig area has not been studied in terms of geochronology. According to general position of Cenozoic volcanics, it is a continuation of the Lakes' Valley area, and corresponding volcanic rocks are probably of the Oligocene–Miocene age.

VIII. Barun-Huray area is the westernmost one, where Cenozoic volcanism of Mongolia is manifested. A small basaltic plateau of the area is composed of thin lava flows, which have not been dated.

IX. Dzabhan area of 50 km^2 is situated in a river valley 20 km away from the settlement Tayshir (Devyatkin, *et al.*, 1973; Devyatkin, 1981; Kepezhinskas, 1979). A volcanic center (stock body and dikes) in the east of the area erupted the stratified lava sequence that is 50 m thick. The sequence includes 3 to 5 lava flows, which are locally intercalated with red beds of the Shand-Gol Formation. The upper flow overlies a member of Miocene pebblestones and sandstones. In distinction from lower weathered and carbonatized lavas, the upper flow with columnar jointing is composed of fresh basalts.

K–Ar dates characterizing the middle (31 ± 2 Ma) and upper (16 ± 0.5 Ma) volcanic horizons have been obtained at the Geological Institute RAS (Devyatkin and Smelov, 1978; Devyatkin, 1981). Accordingly, the lower and middle horizons correspond to the Oligocene, whereas the upper one is of the middle Miocene age (Fig. 3).

X. Hangay area. There are three generations of volcanic rocks in this area.

(1) Thick (dozens of meters) multilayer lava sheets or the so-called watershed basalts occupy the axial part of the Hangay Ridge.

(2) The valley basalts are locally preserved along periphery of the watershed sheets as narrow benches up to 100 m high (e.g., in the Ih-Tamir River valley). Farther downstream, these lava flows are thinner, like, for instance, the basalts near the Ih-Tamir Village, which are 10 to 15 m thick only.

(3) The other valley basalts around the Egin-Daba pass and in the left tributary of the Tuin-Gol River are crowned with lava and cinder cones.

Geochronology is poorly known. One K-Ar date of 28.5 ± 1.0 Ma (early Oligocene) characterizes the middle interval of the watershed lava sequence near the Egin-Daba pass, where the upper flows may correspond to the Miocene. The small valley flows with well-preserved volcanic cones can be definitely attributed to the Quaternary. The watershed lava flows are likely erupted from separate volcanic centers, which cannot be recognized in the relief because of the nival planation at high altitudes.

XI. Gobi Altay area corresponds to crest zone of the Ihe-Bogdo Ridge, where lava flows are as thick as a few tens of meters. It is possible that the area was originally linked with basalts of the Lakes' Valley to be risen up afterward. Geochronology is unknown.

XII. Lakes' Valley area is studied better than others in geological and geochronological aspects. On the South Hangay Plateau, basaltic field covers the gentle slope facing the Lakes' Valley and extends between the Baydarig-Gol and Ongin-Gol rivers for a distance of 300 km, being 50 km wide. The field is related to a series of deep-seated faults of the Bayanhongor zone. Several morphological forms of eruptions are distinguished.

(1) Multilayer sheets of large basaltic plateaus are situated in the Tuin-Gol–Tazin-Gol interfluve. These plateaus composed of 10 and more lava flows are as large as 50–70 km^2 and overlie the bedrocks in the north of the area.

(2) Neck bodies of eruption centers are of a complex structure and large in places (e.g., the Mt. Dushi-Sharo-Ula). The chain of necks extending from the southeast to the northwest outlines a zone of volcanic conduits.

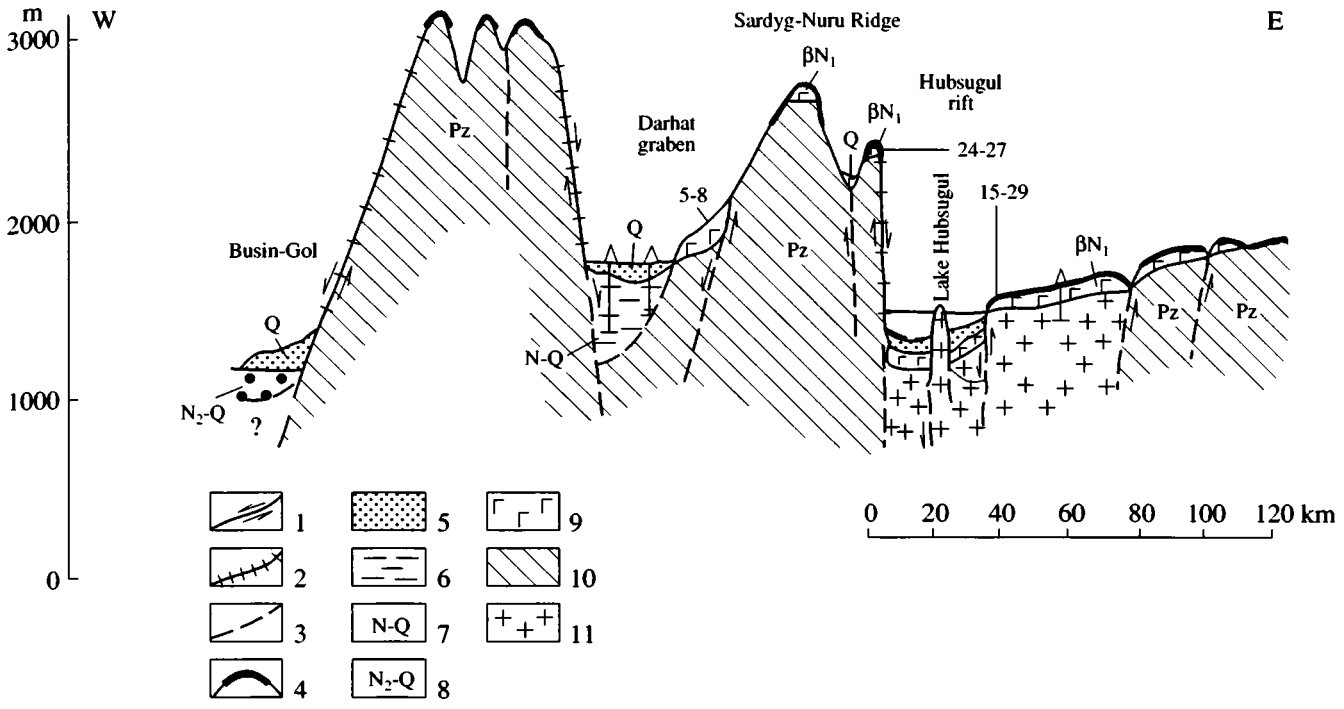

Fig. 2. Basalts of the Hubsugul area and their position in latitudinal profile across rift structures of northern Mongolia (K–Ar age of rocks in Ma): (1) faults, recent and activated in the Neogene–Quaternary time; (2) faults marked by seismic activity; (3) inferable and inactive faults; (4) segments of deformed peneplanated surface; (5) Quaternary (Q), (6) neotectonic and rift (N–Q), (7) Neogene–Quaternary, and (8) Pliocene–Quaternary structural complexes; (9) Neogene basalts; (10) Paleozoic rocks and (11) granitoids of the basement.

Some necks retain their cap of scoriaceous lavas, but simple stocks represent the most common case.

(3) In the northern part of the Lakes' Valley, lava flows are intercalated with the Oligocene–Miocene sedimentary rocks. The interlayering of lavas and sediments has been traced well along a profile connected with the Dushi-Sharo-Ula neck (Devyatkin, 1981). Geochronology of basalts is studied here in detail (Mellet, 1968; Devyatkin *et al.*, 1973, 2002; Gabuniya *et al.*, 1975; Badamgarav *et al.*, 1975; Devyatkin and Smelov, 1979; Daxner-Höck *et al.*, 1997; Höck *et al.*, 1999).

A series of K–Ar dates was obtained in the 1970s–1980s for basalts intercalated with the Oligocene sedimentary rocks. As is established for this part of the sequence, there are two geochronological levels of volcanic rocks within the time span of 32–28 Ma.

The third lava flow overlying the Loo Formation in the Miocene part of the sequence yielded the age values ranging from 20 to 16 Ma (Devyatkin, 1981). The subsequent ^{40}Ar–^{39}Ar dating of 31 samples from three horizons (Daxner-Höck *et al.*, 1997; Höck *et al.*, 1999) specified the following geochronological levels: (1) 30.4–32.1 ± 0.3 to 0.8 Ma (16 dates, lower Oligocene, the Shand-Gol Formation base); (2) 27–29 ± 0.4 to 0.9 Ma (6 dates, the Shand-Gol Formation); (3) 12.2–13.2 ± 0.2 to 0.7 Ma (9 dates, middle Miocene above the Loo Formation).

At present, the last levels are the reference ones for the central part of the Lakes' Valley. Nevertheless, relationships between the Loo Formation and second basaltic horizon remain to be problematic, because fauna from this level is of the Oligocene age. In terms of lithostratigraphy, sediments of the level are closer to the Shand-Gol Formation than to the Loo Formation of the Miocene. In the stratotype and other sections, the persistent erosion surface is a criterion of the Loo Formation base. The Miocene dates ranging from 14.8 to 16 Ma are known for other basaltic sites of the area. It is likely, therefore, that volcanism of the Miocene stage was of a multiphase character in the Lakes' Valley, and corresponding volcanic rocks may be different in age within limits of the stage.

XIII. South Gobi area of a considerable size includes the eastern termination of the Gobi Altay, the Transaltay Gobi, and a southern part of the Lakes' Valley adjacent to the Dalan-Dzadgad Mountains. Volcanic rocks are irregularly studied here. They occur as remnants of lava sheets overlying Paleozoic and Mesozoic rocks and as neck bodies. A sublatitudinal chain of large necks is traceable in the Transaltay part along the southern flank of the Arts-Bogdo Ridge. These necks mark a fault extending for a distance of 100 km. The necks (more than ten in total) are up to 500 m in diameter near the base and crosscut the Lower Cretaceous sedimentary and volcanic rocks, raising above the sur-

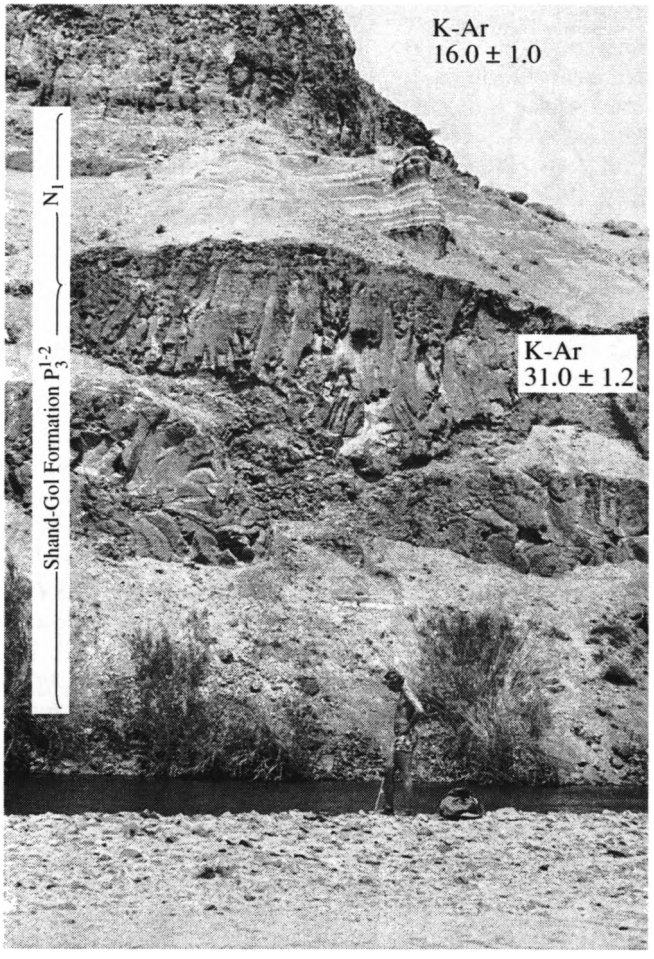

Fig. 3. Dzabhan area: three basaltic lava flows on the right side of the Dzabhan River (K–Ar ages in Ma).

face for 100–150 m. The K–Ar date for the westernmost neck (Mt. Hatan-Sudal) is 57 ± 0.2 Ma (Devyatkin, 1981). Shuvalov and Nikolaeva (1979) studied volcanic rocks in the central part of the area, southeastward of the Dalan-Dzadgad Mountains, where the lava flows overlying the Hashat Formation of Paleocene (?) deposits yielded the K–Ar date of 53 ± 8 Ma, and the underlying basalts are dated back to 61 ± 8 Ma. Both dates are of a low precision and problematic because of absence of biostratigraphic datums.

The date of 47 ± 2.0 Ma characterizes the neck Barun-Tabun-Hayrhan-Ula that intruded the Upper (?) Cretaceous red beds northeastward of the Arts-Bogdo Ridge. Two K-Ar dates of 51 ± 2 Ma (Gabuniya *et al.*, 1975) and of 37 ± 1 Ma (Devyatkin and Smelov, 1979) have been obtained for lava flows overlying the Paleocene Hashato Formation at the Bulgan locality. The Upper Cretaceous volcanics of the South Gobi area do not differ from Cenozoic ones either in composition, or in morphology of their bodies. Thus, Cenozoic volcanism of the area is inherited from the Late Cretaceous volcanic activity, being of the platform type.

XIV–XV. Mandal and Undur-Shili areas. Separate outcrops of volcanic rocks between Mandal and Sain-Shand mountains have been studied and described in detail by Saltykovskiĭ and Genshaft (1985). In both areas, remnants of basaltic sheets form elevations with small smoothed cone-shaped tops (up to 200 m in diameter near the base) or isolated hills up to 100 m high. Columnar jointing characteristic of basalts and morphology of the elevations suggest that they represent individual necks. In opinion of mentioned researchers, ring (5–7 m) and radial dikes observable in the areas under consideration are also remnants of several large volcanic centers, which are now eroded to a high extent.

Three K–Ar dates of 59 ± 5, 59 ± 3, and 68 ± 3 Ma characterize the whole-rock samples of basalts from the Mandal area. The last date indicates presence of Upper Cretaceous lavas. Two dates of 57 ± 3 and 43 ± 3 Ma have been determined for lavas of the Undur-Shili area (Ageeva *et al.*, 1988).

XVI. Dariganga area of about 10000 km^2 is in the southwest of Mongolia. This is the vast basaltic plateau gradually rising up to the altitude of 1400 m. Dozens, if not hundreds, of volcanic edifices of the plateau are different in size (100 to 300 m high), being unequally eroded. The plateau is of a complex inner structure. In the basal portion of volcanic sequence that overlies Paleozoic and Mesozoic deposits, there are intercalations of shingle, quartz sandstone, red and variegated clays. These sediments yield bones of Miocene–Pliocene mammalian fauna (Devyatkin, 1981). The initial volcanic activity in the Dariganga Plateau was likely triggered in the Miocene, when several volcanic centers appeared above the ancient denuded plain and lava flows did not merge to form a single lava field.

The K–Ar date of 18 ± 0 Ma (Genshaft *et al.*, 1990) characterizing the base of the Dariganga lava sequence was obtained for lower lavas of the Meltsog-ele section in the south of the plateau. It means that volcanic activity in the Dariganga Plateau commenced in the middle Miocene to form the first volcanic complex (Boishenko *et al.*, 1990).

The second complex includes stratified lava flows (ten and more), which flooded the plateau surface. Coarse pyroclasts and tuffs outline fields of explosive volcanism. Kepezhinskas (1979) described a sequence of this rocks (30 to 40 m thick) intercalated with red beds, shingle, and boulder horizons. The sequence is intruded by dikes and stock-like bodies of basalts. Similar sequences are characteristic of the Mt. Ashate, Dzun-Nerete, Bayan, and Bayan-Hure localities. Four K–Ar dates obtained for the second complex (Ageeva *et al.*, 1988) range between 3 and 6 Ma, thus corresponding to the uppermost Miocene (probably, the basal Pliocene included). The analyzed rocks have been sampled in the western sites Ihe-Ula, Baru-Narite-Ula, Ashatyn, and Muhuryn-Bulak. The third volcanic complex of the Dariganga Plateau corresponds to scoria-

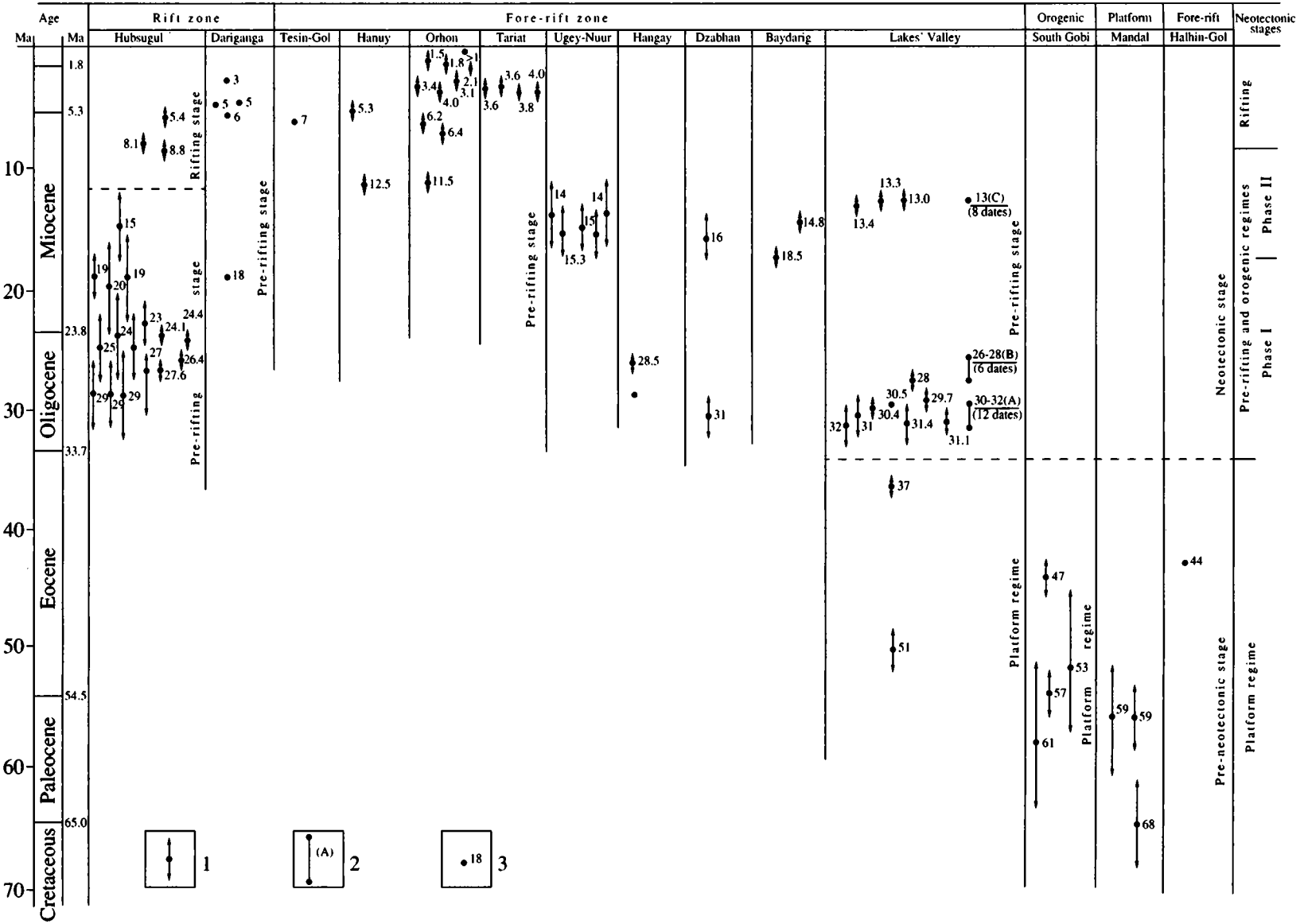

Fig. 4. Geochronology of basalts in different neotectonic zones of Mongolia: (1) K–Ar dates with uncertainty intervals, Ma; (2) ranges of ^{40}Ar–^{39}Ar dates (after Höck *et al.*, 1999) for geochronological horizons A (30–32 Ma), B (26–28 Ma), and C (13 Ma); (3) preliminary dates, Ma.

ceous lavas of well-preserved surficial volcanoes, some of which erupted small lava flows. According to historical records in China, some volcanoes of the plateau demonstrated activity in the 13th and 14th centuries. Rocks of these, undoubtedly Quaternary (Holocene) volcanoes have not been subjected to radiometric dating.

XVII. Halhin-Gol area is poorly studied. It is in the extreme east of Mongolia, where lava flows cover terraces of the Numurgin-Gol River. The lavas are up to 100 m thick. They used to be considered as very young, overlying the Quaternary terrace deposits. This is probably a mistake, because the available K–Ar date of 44 Ma (Genshaft *et al.*, 1990) indicates the much older age of volcanic activity in the area.

The radiometric dates presented above imply the following successive stages of Cenozoic volcanic activity in Mongolia (Fig. 4).

A. Basalts of pre-neotectonic stage (Paleocene–Eocene) are characteristic of southern Mongolia, and corresponding volcanic activity was inherited here from the Late Cretaceous volcanism. The lava fields in question are concentrated in latitudinal depressions located either between the Severey–Nayon and Nomgon–Gurvan–Sayhan ridges, or near the base of the Burun-Tsehe and Arts-Bogdo ridges. The early Oligocene lavas occur among these basalts sporadically. A separate Paleogene volcanic center is situated in the Mandal-Gobi–Undur-Shili region.

B. Basalts of the neotectonic stage (Oligocene–Quaternary) are divisible into several stratigraphic complexes.

(1) The Oligocene complex is characteristic of the Lakes' Valley. It includes two geochronological levels of 28 and 32 Ma, which are identified based on ^{40}Ar–^{39}Ar dates (Höck *et al.*, 1999) and are correlative well with biostratigraphic data (Devyatkin, 1981; Devyatkin *et al.*, 2002). The age range of the complex can be a bit wider. The field of basalts between the Dzabhan River and Lake Ulaan-Nuur is about 450 km long and presumably marks a zone of pre-rifting faulting, which affected the Cretaceous–Paleogene platform.

(2) Basal plateau basalts of the watershed Hangay Ridge, for which a single date of 28 Ma is known, may belong to the same complex.

(3) The Oligocene–early Miocene complex includes lavas widespread in the fore-rift zone, western wall of the Hubsugul rift, and along the southern continuation of this structure. Lavas presumably erupted during two, the Oligocene (29–24 Ma) and early Miocene (24–15 Ma) phases.

(4) The middle Miocene complex (13–18 Ma) is characteristic of the Dashinchilen trough extending eastward from the Lake Ugey-Nuur to the Tola River. Lavas partially filling in the trough are mostly confined to its northern wall.

(5) The late Miocene complex (8–3.5 Ma) includes lavas of the Tesin-Gol, Hanuy, and Orhon rivers in the northern Hangay. Basalts in the southern continuation of the Hubsugul rift (8.4–5.4 Ma) also belong to this complex. They erupted when rifting advanced southward after opening of central and northern parts of the Hubsugul graben. Chains of volcanoes in the Dariganga plateau (6–3 Ma) can be also regarded as manifestations of the pre-rift faulting and associated eruptions. However, the process did not resulted here in formation of the true rift grabens.

(6) The Pliocene–Quaternary complex (3.5 Ma to 10 ka) characterizes two adjacent areas of the northern Hangay: the Orhon River with its left tributaries and the Tariat graben. Volcanic activity was manifested in axial and marginal zones of the corresponding sublatitudinal grabens, which originated in response to the pre-rift faulting. Volcanic activity developed in three separate phases (3–4, 1.8–2.1 Ma, and Quaternary time). Youngest in the complex are rocks of the Quaternary volcanoes Tariat (approximately 6 ka), Togo (Orhon–Selenga interfluve), and Tsagaan-Dzurihe in the Orhon River valley (less than 1 Ma).

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Biogeochemical Study of Lower Sarmatian Bivalves from the Southern Margin of the Ukrainian Shield and Their Stratigraphic Significance

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Abstract—A possibility of using biogeochemical data to define synchronous stratigraphic levels in Sarmatian sections of southern Ukraine is considered. Data on chemical composition of bivalve *Donax dentiger* Eichw., *Ervilia dissita* Eichw., *Obsoletiforma obsoleta* (Eichw.), and *Venerupis vitaliana* (Orb.) shells from three known lower Sarmatian sections of the Ukrainian Shield southern margin are presented. In these sections, intervals with unusual distributions of some chemical elements (Na, Sr, Mn, B, and others) in molluscan shells are established. The data obtained suggest that hydrochemistry in the early Sarmatian brackish-water basin that covered the margin of the East European platform was influenced by collision-related tectonic movements in the Mediterranean mobile belt. The following consequences of tectonic events are inferred from biogeochemical data: (1) isolation and desalination of the Sarmatian basin; (2) activation of fractures in the southeastern margin of the Ukrainian Shield; (3) probability of mud volcanism. The established biogeochemical events are contrasting, steady, and thus can be used to define synchronous stratigraphic levels and to correlate Sarmatian sections.

Key words: stratigraphy, biogeochemistry, bivalves, Sarmatian regional stage, tectogenesis, Eastern Paratethys.

INTRODUCTION

One of the main purposes of applied biogeochemistry is to reveal criteria, which can be used for subdivision of geological sections and for definition of their synchronous stratigraphic levels, although not many examples of successful realization of this task can be cited. In this connection, worth of mentioning is the recent work by Brand and Brenckle (2001) dedicated to chemostratigraphy of the Mississippian–Pennsylvanian boundary in the Arrow Canyon GSSP (Nevada, the United States). The successful biogeochemical study of ancient objects such as Carboniferous brachiopods implies that this method can be also applied to Cenozoic (Vasil'ev, 1997), in particular, to Sarmatian sediments (Vasil'ev, 2002).

The Sarmatian development stage of the Eastern Paratethys was very contrasting and unusual. Precisely this stage characterizes best the essence of the Paratethys that has been formed on the northern margin of the Tethys after isolation by orogenic movements in the Alpine fold belt. It is not unfounded therefore that tectonic movements determined development of the Paratethys, the Sarmatian sea included (Nevesskaya *et al.*, 1986).

Orogeny in the Mediterranean mobile belt played a significant role in evolution of the Eastern Paratethys. During a considerable time span of the Mesozoic and throughout the Cenozoic, geodynamics of the region was determined by collision between the African and Eurasian lithospheric plates, the former of which moved mainly northward (Zonenshain *et al.*, 1990; Shcherba *et al.*, 2001). The active margin of the Eurasian continent with volcanic arcs and back-arc oceanic basins existed south of the East European craton until the Oligocene. In the Oligocene, subduction gave way to collision, which activated nappe thrusting and orogenic processes in the Great Caucasus, Zagros, and other zones along the Arabian promontory, and was responsible for isolation of the Paratethys. The Sarmatian Age was marked by activation of tectonic processes in Asia Minor and Iran, and, since the middle Sarmatian, intense tectonic movements influenced the Greater and Lesser Caucasus.

Variations in the motion rates of the Eurasian and African plates were responsible for intermittent isolation of the Eastern Paratethys from the Tethys, which resulted in rhythmical changes in hydrological systems. Particularly distinct sea-level and salinity fluctuations were in the Sarmatian basin, where transgressive–regressive cycles 1.5–2.0 and 1.0 m.y. long are registered for the early and middle Sarmatian (Chepalyga, 1991). Because of isolation, the Sarmatian basin represented during most of its history a brackish sea with intermittently changing configuration and salinity not exceeding usually 14–15‰.

The lower Sarmatian sediments in Southern Ukraine are represented by highly variable clays, limestones, and sands. Sands and sandy coquinas are of a limited

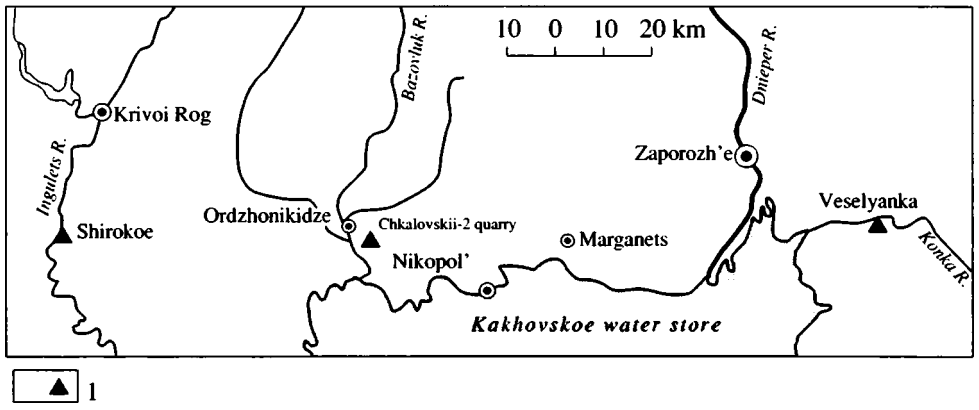

Fig. 1. Localities of studied lower Sarmatian sections (1) in the southern margin of the Ukrainian Shield.

distribution, but precisely these sediments yield the most diverse molluscan assemblages. The lower Sarmatian sandy sediments are distributed, par example, along the southern margin of the Ukrainian Shield, where the well-studied Sarmatian sections of the Eastern Paratethys are localized (Paramonova and Belokrys, 1972). Biogeochemical study of bivalves from these sections provide new information on lower Sarmatian stratigraphy and paleogeography of the southern Ukrainian Shield.

MATERIALS AND METHODS

The lower Sarmatian sections near the Veselyanka Village (Konka River basin, Zaporozh'e oblast), Shirokoe Village (left bank of the Ingulets River, Dnepropetrovsk oblast), and in the Chkalovskii-2 quarry in the vicinity of Ordzhonikidze (Dnepropetrovsk oblast) are examined (Fig. 1). The sections are well studied by traditional methods and described in several publications (Paramonova and Belokrys, 1972; *Stratigraphy...*, 1975; Belokrys, 1976).

Bivalve shells of good preservation and sufficiently wide stratigraphic range were sampled for biogeochemical studies. Shells of *Donax dentiger* Eichw. were collected from the Veselyanka Village section, of *Ervilia dissita* Eichw. and *Obsoletiforma obsoleta* (Eichw.) from the Chkalovskii-2 quarry, and of *Venerupis vitaliana* (Orb.) from exposures near the Shirokoe Village.

Sampling and sample treatment is done in line with traditional procedures (Vasil'ev and Zhuravel', 1988). Chemical composition of shelly matter set in the carbon electrode channel was determined using the emission spectral analysis with the spectrograph STE-1 as registering equipment. The samples preliminarily buffered with NH_4Cl were trice ignited in the carbon channel under the electric arc of the alternating 20A current. The check samples were studied, using the ICP method and the ICP-AES equipment designed by the Therno Jarrell Ash Corporation. Concentrations of Mg, Sr, Na, Cu, Ti, Mn, Zr, B, Al, Si, Ce, La, Y, and other elements were determined.

By the bed-to-bed sampling, relatively thick beds were subdivided into intervals with due account for their lithological peculiarities and distribution of fossils. Each bed or interval is characterized by 5 to 7 samples with subsequent averaging of analytical results. Student and Fischer criteria were used for statistic comparison of the results.

CHARACTERISTICS OF STUDY SECTIONS

N. P. Barbott de Marnie was first to describe and subdivide the complete section of Sarmatian sediments in 1866. Since then, the *Sarmatian Stage* became a subject of booming discussion among experts in Cenozoic stratigraphy who emphasized significant problems in correlation of relevant sections (Paramonova and Belokrys, 1972).

The Sarmatian regional stage is subdivided into three regional substages, which reflect different development phases of the basin: the early, middle, and late (Andrusov, 1961). In the early Sarmatian, when the basin lost its connection with the World Ocean, water salinity sharply decreased, almost all stenohaline bivalve species died out and remained euryhaline forms experienced some changes. The middle Sarmatian was marked by appearance of many new species, of which only representatives of the *Mactra* genus survived in the late Sarmatian (Nevesskaya *et al.*, 1986).

Many researchers divide the lower Sarmatian regional substage in two parts, which are readily recognizable based on the bivalve fauna (Didkovskii *et al.*, 1970). The basal layers of the Sarmatian regional stage are probably exposed in the Konka River basin, where limy–sandy sediments with the characteristic Sarmatian fauna rest immediately upon the Konkian regional stage.

In the Veselyanka Village outskirts, lower Sarmatian quartz sands are mostly fine-grained, yellowish to greenish gray, sometimes grading into compact

coquinas composed of debris and intact shells of small bivalves and gastropods. The bedding is emphasized by irregular distribution of shells, accumulations of which form lenses, isometric nests, and thin laminae sometimes. In the upper part of the section, the number of compact shelly interbeds increases. Bivalve remains are also abundant in sands, where prevalent are *Mactra eichwaldi* Lask., *Ervilia dissita* Eichw., *Donax dentiger* Eichw., *Venerupis vitaliana* (Orb.), *Plicatiforma praeplicatum* Hilb., *Obsoletiforma ruthenicum* (Hilb.), *Abra reflexa* (Eichw.), and others. Shells of these species are usually fragile except for those of *D. dentiger*, which are well preserved. In the exposed section that is over 5 m thick, faunal remains are irregularly distributed, although the last species occurs through the entire sequence.

In the lower Sarmatian section of the Chkalovskii-2 quarry, sandy and coquina–detrital sediments with rare dark gray coaly–clayey interbeds are also dominant. Light green and greenish gray quartz sands prevail in the lower part of the study sequence. Higher in the section, they grade into coquina–sandy gray-colored limestones. Among bivalves, prevalent in these sediments are *Ervilia dissita* Eichw., *Obsoletiforma obsoleta* (Eichw.), *Plicatiforma plicata* (Eichw.), *Donax dentiger* Eichw., and others. *Mactra eichwaldi* Lask. occurs only in the lower part of the section. The occurrence of *O. obsoleta* and *P. plicata* indicate a relatively high stratigraphic level of the section under consideration, which corresponds to the uppermost lower and upper part of the lower Sarmatian Substage. The integral thickness of the section exceeds 7 m.

In the southern outskirts of the Shirokoe Village, the lower Sarmatian sediments overly Paleogene clays and are composed of light gray quartz sands and coquina–detrital limestones. The bivalve assemblage includes dominant *Venerupis vitaliana* (Orb.), *Ervilia dissita* Eichw., *Mactra eichwaldi* Lask., *Obsoletiforma obsoleta* (Eichw.), and other forms. The thickness of the lower Sarmatian sediments in this section exceeds slightly 2 m, thus being significantly lower than in two other examined sections. Nevertheless, it is probable that the section under consideration corresponds in its stratigraphic range to the section described in the Ordzhonikidze area. One can assume that the studied exposures characterize a complete or almost complete lower Sarmatian section in the southern periphery of the Ukrainian Shield.

RESULTS AND DISCUSSION

Through the lower Sarmatian section in the vicinity of the Veselyanka Village, distribution of microelements in *D. dentiger* shells is contrasting, and most significant variations are characteristic of Na, Sr, B, Ti, Mg, and Mn. Particularly noteworthy is the contrasting distribution of La and Ce, which are usually distributed in molluscan shells uniformly.

The maximum concentrations of nearly all elements mentioned above are characteristic of shells from the relatively narrow interval corresponding to the upper part of Bed 7 and Bed 8 in the upper part of the section (Fig. 2). In this interval, Na and Sr concentrations are minimal, whereas other elements are highly concentrated. Sodium and strontium in aragonite molluscan shells are considered as indicators of water salinity (Borisenko and Vasil'ev, 1990), and synchronous decrease in their contents in *D. dentiger* shells suggests most likely the basin desalination.

Another group of microelements (Mn, La, Ce) is probably of endogenic origin and their elevated contents can be related to activation of fracturing. In the immediate proximity to the study section, there is the Chernigovka structural–geochemical zone, typomorphic chemical elements of which are Nb, La, Ce, Sr, P, Mn, and Mo (Galitskiĭ *et al.*, 1984). Of these elements, Nb, P, and Mo were not determined in examined material, Sr in shells is controlled by organism exchange reactions, and remaining La, Ce, and Mn produce well manifested concentration maximums within that "anomaly". It can be assumed that enrichment of waters and, correspondingly, of molluscan tissues with these elements is related to their submarine influx with deep fluids from the activated faults in the Chernigovka structural–geochemical zone.

Tectonic movements activated the river run-off from elevated areas of the Ukrainian Shield. This was probably responsible for the enrichment of Sarmatian sediments with Ti, as the Middle Dnieper area of Ti–Zr placers stretches in the form of a wide band along the northern shoreline of the Sarmatian basin. The joint occurrence of leucoxene and ilmenite in placers implies hydrologic and biogenic migration of titanium.

The defined geochemical phenomena are probably a consequence of tectonic processes in the southerly-located area of the Alpine collision belt. Orogenic movements of that time were intense and, probably, rhythmical, producing different-scale dislocations that influenced areas adjacent to the Alpine belt in the north, the rigid blocks of the Ukrainian Shield basement included. Accordingly, it is possible to assume that the biogeochemical effect mentioned above evidences one of the rhythms of tectogenesis activation in the Mediterranean mobile belts.

If this is the case, the occurrence of boron among typomorphic microelements is readily explainable. Its excessive quantities are released into environment owing to both the subduction-related magmatism and mud volcanism. In our case, the second source is preferable, because there is an opinion that diapirism and mud volcanism, which accompany collision, are the results of intense tectonic load under strong lateral compression (Limonov *et al.*, 1996). Mud volcanoes were active in the Kerchenskii and Tamanskii peninsulas during the Sarmatian Age and later (Shnyukov *et al.*, 1971), and they could supply boron. Borates are of a

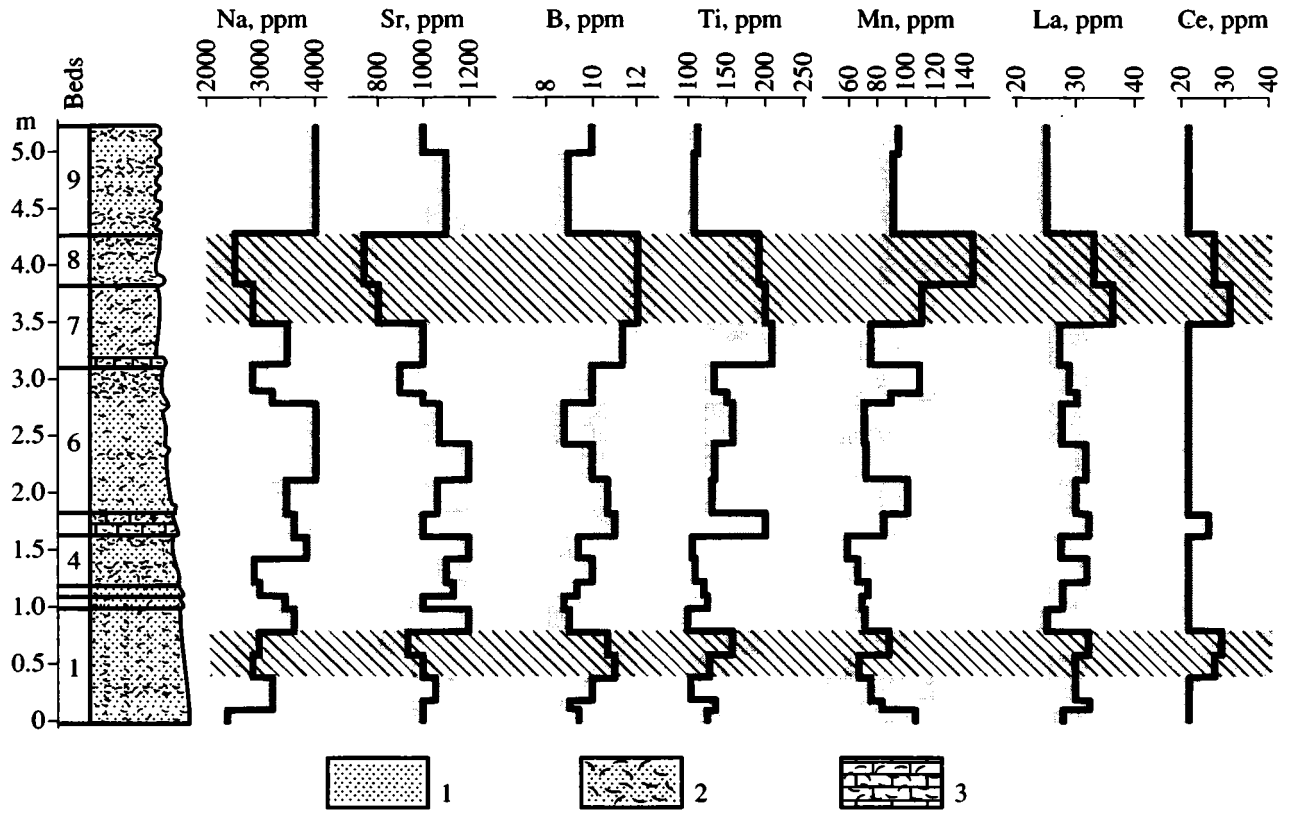

Fig. 2. Distribution of some microelements in shells of bivalve species *Donax dentiger* from the lower Sarmatian sediments of the Konka River basin, outskirts of the Veselyanka Village, Zaporozh'e oblast (thick solid line designates average contents of microelements averaged for 5–7 samples and gray-colored intervals show the uncertainty range): (1) quartz sands with admixture of shelly debris; (2) organogenic–detrital limestones with sand admixture; (3) compact coquina.

high geochemical mobility in low-alkali oxidizing seawater environments, high boron concentrations are established in molluscan shells at a significant distance from its source (Vasil'ev, 1997). This implies that mud volcanism outbursts can be synchronous with tectonic activation periods. The defined interval with a peculiar proportion of microelements in shells of *Donax dentiger* is not the only one. A similar, though less obvious situation is characteristic also of the middle part of Bed 1 (Fig. 2).

The study of *Ervilia dissita* shells from outskirts of Ordzhonikidze (Chkaolvskii-2 quarry) revealed similar biogeochemical phenomena. Sediments of this area includes as well an interval with minimal Na and Sr and elevated B, Pb, Ti, Cu, and Zr contents in shells (Fig. 3). Like in the Veselyanka section, informative microelements are Na, Sr, Ti, and B, and their distribution in shells was likely controlled by the same factors: by the basin desalination that decreased the Na content, by mud volcanism activity in the south that increased B concentration, and by elevated river runoff rising Ti content. Other elements reflect probably a local geochemical situation, the chalcophile specialization of the Kamensk fault (Nechaev *et al.*, 1986).

In addition to the apparent similarity of geochemical phenomena in two study sections, they are different to some extent. In the Chkalovskii-2 Sarmatian section, the interval with *Ervilia dissita* shells enriched in Pb, Cu, and Zr is long that is untypical of the Veselyanka section. It is likely that the intense removal of men-

Dependence of elevated microelement concentrations in molluscan shells on geochemical specialization of adjacent structural–geochemical zones (Galitskiĭ *et al.*, 1984; Nechaev *et al.*, 1986)

Locality	Association of microelements in shells	Adjacent structural–geochemical zones and their geochemical specialization
Veselyanka Village	Mn, La, Ce	Chernigovka structural–geochemical zone (Nb, La, Ce, Sr, P. Mn, Mo)
Shirokoe Village	Mn, Cr, Ni	Ingulets–Krivoi Rog geochemical zone (Mo, Cr. Ni, Sn, Yb, Cu, Mn)
Chkalovskii-2 quarry	Pb, Cu, Zr	Kamensk fault (Mo, Cu, Pb, W, Zr)

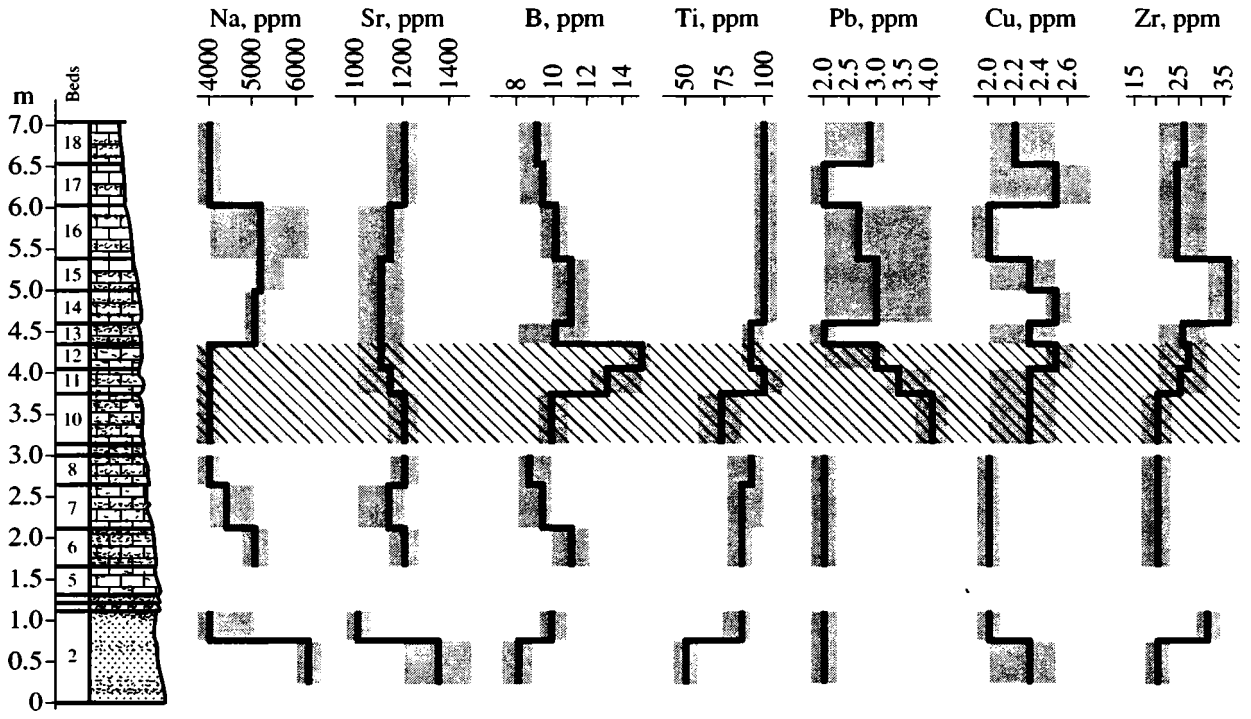

Fig. 3. Distribution of some microelements in shells of bivalve species *Ervilia dissita* from the lower Sarmatian sediments of the Chkalovskii-2 quarry, outskirts of Ordzhonikidze, Dnepropetrovsk oblast (symbols as in Fig. 2).

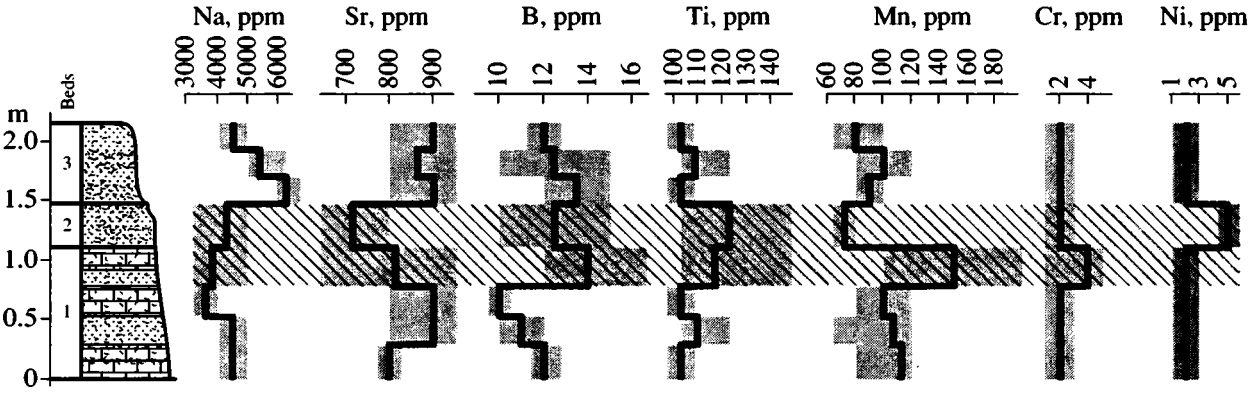

Fig. 4. Distribution of some microelements in shells of bivalve species *Venerupis vitaliana* from the lower Sarmatian sediments in the Shirokoe Village outskirts, Dnepropetrovsk oblast (symbols as in Fig. 2).

tioned elements from mineralization zones related to tectonic movements was long lasting and continued after their termination, which can be explained by the peculiar structure of fluid feeding channels. Figure 3 demonstrates clearly that shells from the lower part of the section are depleted in Pb, Cu, and Zr as compared with those from its upper part.

The lower Sarmatian section near the Shirokoe Village is incomplete (Fig. 4). Nevertheless, it encloses an interval reflecting tectogenesis. The uppermost part of Bed 1 and Bed 2 yield shells of *Venerupis vitaliana* with the minimal Na and Sr and elevated B, Mn, Cr, and Ti concentrations. The typical elements occurring throughout the section are Na, Sr, Ti, and B, while a new geochemical association of Mn, Cr, and Ni is likely related in origin to a submarine discharge of local mineralization zones. Moreover, a peculiar zoning is distinguishable: the elevated Mn and Cr contents are confined to the lower part of the indicated interval, whereas Ni is more concentrated in the upper one. These siderophile elements are undoubtedly typomorphic and characteristic of the Ingulets–Krivoi Rog geochemical zone (Galitskiĭ *et al.*, 1984).

The defined geochemical effects recorded in bivalve shells are almost identical in the studied lower Sarmatian sections. Distinctions are connected only with ele-

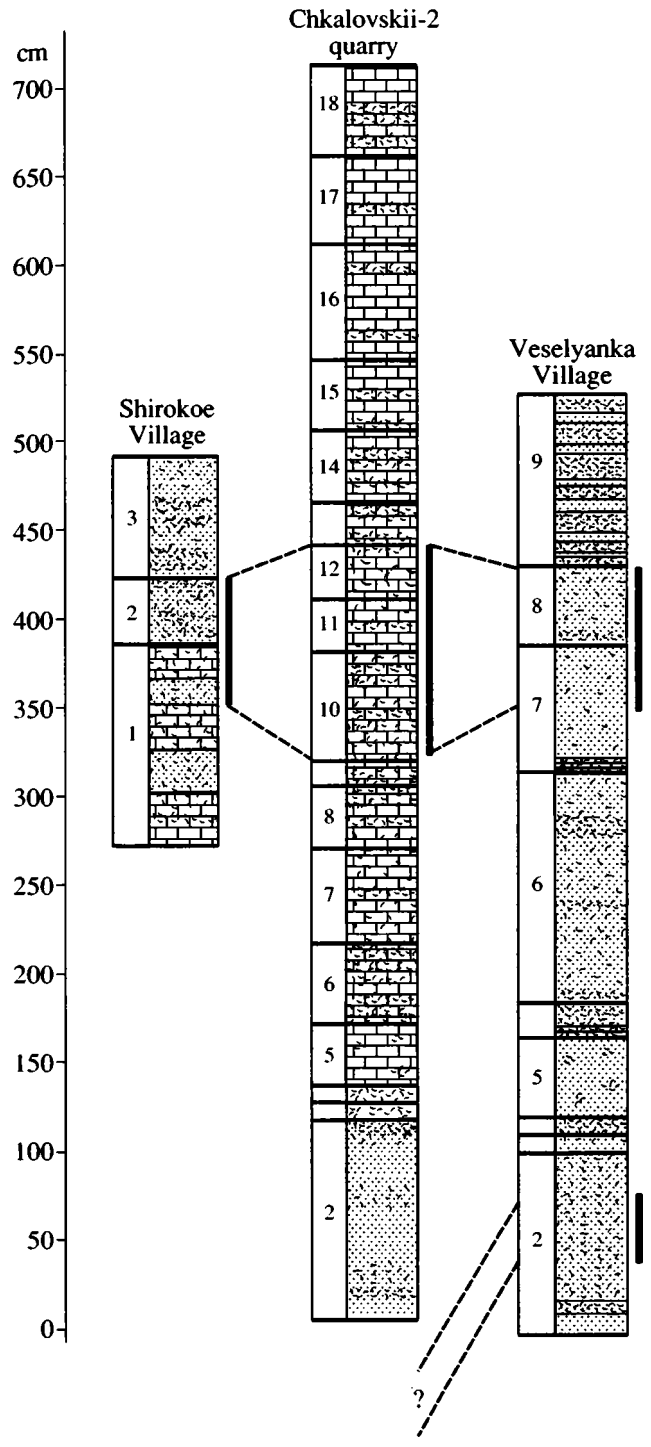

Fig. 5. Correlation of the investigated lower Sarmatian sections based on biogeochemical data (symbols as in Fig. 2).

ments of local origin. Thus, there were likely only two large-scale tectonic events during the Sarmatian development stage of the Eastern Paratethys, which were related to collision processes in the Mediterranean mobile belt. It should be noted that, despite the different extent of manifestation (the earlier event was weaker than the later one), they resulted in similar consequences for the basin evolution (Fig. 2). Based on established biogeochemical parameters, the studied lower Sarmatian sections can be correlated between each other (Fig. 5).

Of particular interest are also associations of microelements carried by ascending fluids. Tectonic movements induced by collision in the south activated the mass transfer in geodynamic areas of active faulting. The geochemical spectrum of submarine associations is very diverse and includes elements with different chemical properties and of different mobility (table). These are ions with the low activity in seawater as a rule, and therefore their migration distance is insignificant and dispersion areas are local.

CONCLUSION

The data obtained lead to the following conclusion:

(1) In the early Sarmatian, hydrochemistry of a semimarine basin that covered the margin of the East European platform was substantially influenced by tectonic processes in the southerly-located area of the Alpine collision belt.

(2) The biogeochemical data suggest that orogeny in the Mediterranean mobile belt was responsible for isolation and desalination of the Sarmatian basin, intensification of the river runoff from northern slopes of the Ukrainian Shield, and, probably, for activation of mud volcanism in the Kerch–Taman region.

(3) Based on biogeochemical data, local geochemical phenomena are also well recognizable in addition to the above regional processes. In my opinion, they are related to the geochemical influx from thermodynamically active fractures in the basement during the epoch of tectonic activation. The spectrum of chemical elements concentrated in seawater was determined in general by geochemical specialization of structural–geochemical zones located close to study areas. The substance migration was realized through a submarine discharge,

(4) The established biogeochemical phenomena are contrasting, easily recognizable, and can be used as indicators of synchronous stratigraphic levels.

Reviewers I. A. Goncharova and I. G. Shcherba

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Critical Remarks to the Paper by S.A. Laukhin, A. Ronen, V.A. Ranov, G.A. Pospelova, Ya.M. Burdukevich, Z.V. Sharonova, V.A. Volgina, O.A. Kulikov, V.K. Vlasov, and A. Tsatskin *New Data on the Paleolithic Geochronology in Southern Levant*

Ya. V. Kuz'min

Received June 25, 2001

Journal *Stratigraphy and Geological Correlation* (2000, vol. 8, no. 5, pp. 82-95) published the paper of S. A. Laukhin and coworkers "New Data on the Paleolithic Geochronology in Southern Levant" that contains RTL dates for sediments bearing artifacts in certain regions of Israel. After a close inspection of the paper, I doubt the validity of published RTL dates and reliability of the authors' conclusions about the age of the Yeron locality, the most ancient Paleolithic site in Levant and entire Eurasia (if the age of sediments is determined correctly). I believe that my remarks will be useful not only for readers, but also for authors of that paper.

The first remark concerns the inversion of the RTL dates that is repeated in several sections. For instance, in the Yeron section the age measurement of 900 ± 230 ka (RTL-720) was obtained for the sample taken 35–40 cm *above* that dated at 430 ± 110 ka (RTL-719). For the sediments overlain by the basalts with the K–Ar age of about 2.5 Ma, two age determinations were obtained in addition to the date of 2800 ± 700 ka (sample RTL-717 from the depth of 25 m). One corresponds to 1270 ± 300 ka (RTL-716, the depth of about 5 m) and the other one to 1300 ± 330 ka (RTL-718, the depth of about 23 m). In the Habonim section, the date of 130 ± 33 ka (RTL-721) is defined *above* the date of 107 ± 27 ka (RTL-722). The authors do not explain reasonably that discrepancy between the obtained ages and stratigraphy of the sections, which is especially considerable for the "key" Yeron locality.

The second, most essential remark is concerning the date of 2800 ± 700 ka obtained for the lower part of sediments occurring below basalts in the Yeron section, which contain artifacts. On page 501 the authors write: "...one sample taken from the lens of rough pebbles at the depth of 25 m yielded the value of 2800 ± 700 ka (RTL-717), which may illustrate *the real age* [*italicization* is mine] of red clay." The discrepancy between RTL-716, RTL-718, and RTL-717 is interpreted on that page in the following way: "The dates obtained for red clays above the depth level of 25 m are rejuvenated under influence of some physical factors. This issue is an object for discussion of physicists, the experts in isotopic and radiothermoluminescence methods, and we do not consider it here."

This standpoint is a reason of perplexity. The authors announce that the readers of the journal, mainly geologists, will understand nothing in "physical factors" responsible for the complete inconsistency of the obtained dates, which create the basis for issues discussed in the paper. In such case, the paper should be published in a physical but not geological journal! This has not been done, and the reason is likely as follows.

As is known, the lower age limit for determination by the luminescent methods is about 800 ka (Aitken, 1997, pp. 209–211). Nevertheless, the authors repeatedly obtained the dates "close to the real age" of up to 3 Ma in contradictions with other approaches to the luminescent dating. As for the RTL method proper, the considerable doubts upon reliability of its principles were cast by experts in isotopic and thermoluminescence methods (Huntley, 1992), and these doubts have not been clarified by Vlasov and Kulikov, the authors of the RTL method and co-authors of the paper under discussion. In terms of international scientific standards, this means that both the RTL method and its results cannot be considered as correct at present.

It should be emphasized that the RTL dates are not reproducible by other luminescent methods. For instance, the RTL dates for sediments of the Diring-Yuryakh site in Yakutia are as follows: over 1800 ka (RTL-454) below the cultural layer, 2900 ± 900 ka (RTL-424), and over 1100 ka (RTL-453) above the cultural layer (Mochanov *et al.*, 1992, p. 131). The thermoluminescence method yielded later the substantially younger ages for the same sediments: 366 ± 32 ka (OTL-472) below the cultural layer and 267 ± 24 ka (OTL-471) above the cultural layer (Waters, 1997, 1999).

Therefore, the physical basics of the RTL method is still unproved, and the RTL dates cannot be reproduced by other luminescent methods and contradict the ^{14}C age determinations for the same sediments, for example in the Altai Mountains (Kuz'min, 2000, p. 35). This sig-

nifies nothing else but the fact that *the age determinations in the paper under discussion, in particular those obtained for the Yeron locality, cannot be regarded seriously until the reliability of the RTL method is additionally substantiated.*

As regards the authors' reluctance to discuss the reasons of the age rejuvenation, it should be noted that another academic geological journal "Geology and Geophysics" consider both the results and principles of the radiothermoluminescence method of age measurement (Perevalov and Rezanov, 1997). I hope that editorial board of "Stratigraphy and Geological Correlation" will be more meticulous for papers of this kind.

Reviewers M.N. Alekseev and E.V. Devyatkin

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Reply to the Critical Review by Ya.V. Kuz'min

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The review creates an impression that the article by Laukhin *et al.* (2000) is dedicated to RTL dating. This is far from being the case. The article is based on geological and stratigraphic data. The RTL dates themselves solve nothing in our work. The oldest artifacts at the Yeron site occur in clays overlain by basalts dated by the K–Ar method back to 2.51 Ma. Two levels of the basalts in the western wall of the Jordan rift are studied in detail and reliably dated. The Jordan rift was under investigation for a long time and is among the most comprehensively studied continental rifts. The Yeron site is connected with one of the basalt levels in this rift, which was dated (2.4 to 2.5 Ma) long before the first RTL dates were obtained (Ronen, 1991). The Habonim section is related to hamra (this and other local names are explained in (Laukhin *et al.*, 2000), which is traceable throughout the entire Maritime Plain being dated in many places (Korobkov, 1978 and others). The RTL dates obtained only specify geochronology of kurkars. To test reproducibility of RTL and TL dates obtained at the H. and G. Valladas laboratory, we dated Bed E of the Tabun Cave. It appeared that accuracy of measurements is high. At the Diring-Yuryakh site (Yakutia), which is mentioned in the review, the oldest TRL dates are obtained for red beds underlying sands with a cultural layer. The date of approximately 1000 ka was obtained for the cultural layer not by the RTL method (only?), but using also the paleomagnetic method and microtheriofauna, although the latter is impoverished (Alekseev *et al.*, 1990). Almost a quarter of the review is dedicated to the Diring-Yuryakh site, which is even not mentioned in our article. What has southern Levant to do with that site? Finally, the Bizat Ruhama site is dated by the paleomagnetic method. The paleomagnetic studies were carried out in different years and at different laboratories (at the Geological Institute RAS and Institute of the Earth Physics RAS) and provided similar results. The paleontological data (Horowitz, 1979 and others) are consistent with these dates. It is not our fault that the RTL dating confidence appeared to be compatible with paleomagnetic data. In this connection, the reviewer's attempt to convince the reader that the article deals with the RTL dates and is based only on them looks biased.

What is our article about? The excellent exposures, wonderful roads, and perfect organizing work of scientists from the Haifa University made it possible to collect during two field expeditions so representative material that its processing took over five years and resulted in over 10 articles, almost all of which are published abroad. We cannot mention all these publications because of the limited volume of the article and note just most important works of Ronen *et al.* (1998, 1999, 2000) and Laukhin *et al.* (2001). We decided to present our results also in Russia because we thought it is unethical to publish them only abroad, especially taking into consideration the limited scientific information from there during the last 15 years.

What are the real shortages of our paper? It deals only with problems of geochronology and Paleolithic geology, whereas similarly important materials on archeology and paleopedology are almost completely ignored. Moreover, the geochronological data are also presented with insufficient thoroughness. Nevertheless, this is the first Russian article dedicated to data on very old (Matuyama Chron) Bizat Ruhama site discovered by one of the authors. This site is interesting in both the geological and archeological respect. The article considers, for the first time in the Russian scientific literature, the data on the peculiar Mousterian Habonim site connected with marsh sediments (we were first to discuss this issue abroad as well). We studied and first dated the lower cultural layers in the worldwide known Tabun site. There is no need to enumerate all new facts mentioned in the article and ignored by the reviewer: everybody who is interested can read the work.

What the reviewer's censorious remarks are reduced to?

Inversions of RTL dates are presented in such a way as if it is something uncommon. In fact, inversions of radiocarbon dates are usual and virtually unavoidable in sections, from where 15 to 20 samples are dated (Kostenki, Chany, Mt. Afontova 2, and others). The fact that the ^{14}C-dating method is well elaborated and causes of inversions are always obvious is a different matter. The thermoluminescence methods and even their principles are worked out to a lesser extent. This is however quite natural: during the 1950–1960s, inversions of ^{14}C dates seemed frequently also strange. Even now it is difficult to explain inversions in the interval of 12 to 7 ka. In addition, inversions are commonly within

the confidence interval, i.e., they are seeming inversions.

Chinese scientists publish regularly the dates of approximately 3 Ma, but "co-authors of the article" have nothing to do with this. In this connection, the reviewer's statement that the Yeron site is the oldest one in Eurasia is amazing. There are others of similar age. The PS-55 site in Pakistan is dated at 2.47 Ma, and this date is not questioned so far. There are also other examples.

The reviewer "knows that the age of approximately 800 ka represents the lower limit for the luminescence methods" (Aitken, 1997). Meanwhile, M. Frechen, an expert respected in the scientific community and by myself, obtains reliable dates as old as 150–200 ka. Does this mean a chaos in TL dating? Not at all, just different laboratories use different procedures of dating. Dates obtained by V.K. Vlasov can be substantially old, but they have a very wide confidence interval, whereas those obtained by M. Frechen are limited by the terminal Middle Pleistocene, being more accurate and characterized by a narrower confidence interval. Dates obtained for Hellenic amphorae have the narrowest confidence interval. For some reason, the reviewer does not mention this fact.

Nobody argues that physical principles of the TL (RTL) dating are insufficiently worked out. Their further elaboration is desirable, and physicists and experts in isotope chemistry are carrying out this work. They can answer difficult questions, but not geologists or archeologists who apply the methods. Of course, it would be better to use only the best-developed methods. Unfortunately, nothing else, except for the TL (RTL), methods is available for the time interval unsuitable for the radiocarbon dating (50–1000 ka). Some other methods require specific material for dating, which is occasional in sections (thick buried peat, stalactites, and others). For instance, the INTAS project 01-0675 demonstrated that entire West Siberia (one third of Europe!) hosts only eight peat localities suitable for dating by the U–Th method. Other methods are elaborated even lesser than TL methods and their reproducibility is very low, for instance, that of the EPR method (Jelinek, 1973). Therefore, various modifications of the TL dating method are intensely applied worldwide, Russia inclusive, despite the insufficient development of its methodical principles and (frequently) the ambiguous results. Physicists and experts in isotope chemistry are those specialists who should discuss the principles of this method. Any geologist, like me, can discuss only consistency (inconsistency) of TL dates with geological material. The RTL dates mentioned in our article are consistent with both geological facts and data obtained by other independent methods. From this side, the RTL dates discussed in the article seem reliable.

Reviewer M.N. Alekseev

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